MILLENNIAL TO ANNUAL SCALE PALEOCLIMATIC CHANGE

IN CENTRAL ALASKA DURING THE LATE QUATERNARY INTERPRETED

FROM LAKE SEDIMENTS AND TREE RINGS.

A

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By

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Abstract

The theme of this dissertation is the importance of effective moisture (precipitation minus evaporation) in subarctic ecosystems. Interior Alaska has a relatively dry climate with annual precipitation ranging from 25-45 cm. Records from interior Alaska lake sediment cores show low lake levels following the Last Glacial Maximum, with significant increases at 12,000 and 9,000 ^{14}C years B.P. Using lake-level reconstructions and models based on modern hydrologic and meteorologic data, we infer precipitation of 35-75% less than modern at 12,000 yr. BP, 25-45% less than modern at 9,000 yr. BP, and 10-20% less than modern at 6,000 yr. BP. Trees were scarce on the interior Alaskan landscape during the late Pleistocene with birch species appearing about 12,000 BP and spruce species approximately 3500 years later. The correspondence between lake-level and vegetation changes suggests that moisture may have been one of the limiting factors in the establishment of these tree species.

Alaska climate records show a climatic regime shift in the mid-1970s. Less effective moisture is available over the past 30 years because summer temperatures in interior Alaska have been increasing without a concurrent increase in precipitation. Radial growth of white spruce at 20 low elevation stands in interior Alaska declined corresponding with this climatic change. The observation that moisture limits spruce growth in Alaska today is consistent with our inference of moisture limitation in the early Holocene. A 200-year reconstruction was developed based on two tree ring proxies, ^{13}C discrimination and maximum latewood density, which together show excellent agreement with the recorded Fairbanks average May through August temperatures. The first half of the 20th century is characterized by the coolest summers of the 200 year period of reconstruction, while the latter part of the 20th century, particularly from 1974 onward, is characterized by some of the warmest summers of the 200 year period. Mid-19th summer temperatures reconstruct to be as warm as the latter part of the 20th century, which is inconsistent with reconstructions of other regions. It seems likely, based on current information, that these inconsistencies may be real and may reflect regional synoptic conditions unique to interior Alaska. Distinctive decadal scale regimes were identified throughout the record.

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Chapter 1
Introduction and Overview

Growing concern that climate is changing at unprecedented rates due to human activities prevails. Media reports indicate increased occurrences of flooding, droughts, hurricanes, and wildly fluctuating weather conditions worldwide. Global temperatures have increased throughout the 20th century, although not uniformly over the planet. Some researchers have concluded that recent temperatures are of unprecedented warmth relative the past 1000 years (Briffa *et al.*, 2001; Crowley, 2000; Mann *et al.*, 1999). It is well documented that atmospheric carbon dioxide and methane concentrations are increasing and modern levels are at the highest of the past 420,000 years (Petit *et al.*, 1999). Several studies have indicated that natural climate variability cannot account for all of the warming documented during the 20th century and that anthropogenic factors which contribute to greenhouse gas emissions play a large role (Houghton *et al.*, 1996; Overpeck *et al.*, 1997). It is essential to understand how much of the recent climatic change is due to natural variability and how much of it is due to anthropogenic effects. In order to understand natural variability, we must reconstruct past climates. General circulation models consistently predict that the greatest change will occur at high latitude regions. Therefore Alaska is an ideal location to look for these changes and to identify past climate variability.

Climate History of interior Alaska

In order to determine how future climate change may affect this region, we can look back in time at ecological responses to past climate change. Following the peak of the last ice age about 20,000 years ago, temperatures started warming and ice sheets began to melt. Interior Alaska was ice free due to the lack of available moisture. There were extensive glaciers in the Brooks and Alaska Ranges and the Cordilleran ice sheet covered Alaska south of the Alaska Range. By about 11,000 years ago, sea level was approaching that of the present day and most of the glaciers formed during the ice age were greatly reduced.

Summer solar insolation was at a maximum about 10,000 years ago due to changes in the earth's orbital characteristics (COHMAP, 1988). There are indications that it was warmer than present in Alaska at this time, and that some species of deciduous trees may have reached higher latitudes than at present (Hopkins *et al.*, 1981). White spruce did not arrive in interior Alaska until about 8,500 ^{14}C yr. B.P. Results from the original NSF-funded PALE initiative (**P**aleoclimates from **A**rctic **L**akes and **E**stuaries) and this dissertation suggest its arrival was probably delayed beyond the onset of suitable temperatures by drought limitations. All the elements of the modern day Boreal forest were present 6000 yr. BP. Since then, the relative abundance of plant species has shifted while some species have extended their distribution (Ager, 1982).

Neoglaciation began 6000 to 3000 BP producing cooler temperatures and increased storminess (Mason and Begét, 1991; Wiles and Calkin, 1990). Around 1000 AD, a warm period labeled elsewhere as the Medieval Warm Period (Hughes and Diaz, 1994) occurred, inducing a moderation in climate and storminess for several hundred years. Between 1450 and 1890 AD, an interval of relative cold, the Little Ice Age (Bradley and Jones, 1993; Overpeck *et al.*, 1997; Wiles and Calkin, 1990), was identified. This cool period lasted until the mid to late 19th century and was broken up by periods of warmth in the 16th and 18th centuries. Climate started warming in some regions about 150 years ago (Mann *et al.*, 1999), interspersed with periods of cooler summers, and there has been intensified warming over the past 3 decades. This warming in interior Alaska came without a concurrent increase in precipitation, causing a decrease in effective moisture (precipitation minus evapotranspiration) (Chapter 4).

Dissertation Overview

The research presented in this dissertation demonstrates that effective moisture is an important control on lake-levels and white spruce distribution in interior Alaska today, and in the past. This thesis uses data from lakes and tree-rings to serve as proxies for elements of climate change and white spruce tree growth, first by determining modern calibrations, and then by hindcasting to look at past conditions. These tree-ring and lake sediment archives provide information on a range of temporal scales. Lake sediments typically record on scales of tens to thousands of years, allowing resolution of 1 to 100 years depending on depositional processes, while tree-ring records provide annual resolution over centuries. In this dissertation, I use lake-sediment proxies for reconstructing climate at low resolution scales over the past 13,000 years and tree-ring data to reconstruct climate at high-resolution over the past 200 years.

The hydrologic system of interior Alaska is very complex and interesting, due to the large seasonal range in climate and the huge quantities of water locked up in glaciers and permafrost, combined with semi-arid climatic conditions. Precipitation varies widely in Alaska, with high precipitation in the southern coastal mountains, but low annual precipitation in interior basins. Interior Alaska lies within the boreal forest vegetation zone. Large parts of the Alaskan boreal region are currently at the dry end of the continuum of environmental conditions in which the boreal forest exists. Future climatic scenarios for this region vary widely but indicate significant changes. Understanding the response of the boreal forest to past climatic change can enhance predictions of the future of the boreal forest in this region.

There are a large number of lakes, including closed-basin lakes, in interior Alaska, which are sensitive to changing precipitation/temperature regimes. The balance between evaporation, precipitation and groundwater primarily controls the size and water quality of these lakes. Combining modern water budgets with past lake-level history allowed for quantitative reconstruction of past precipitation. Birch and Jan lakes were chosen for this study as part of the PALE study. Their hydrologic characteristics suggest they are sensitive to changes in precipitation. Lake-level changes, determined from sediment core transects, show much lower levels in the past. In Chapter 2, three snapshots in time were modeled (12.5 K, 9 K, 6K) to determine past precipitation. At 12.5 K, model results show that precipitation could have been up to 45% to 75% less than modern, at 9K 25% to 45% less than modern, and at 6K 10% to 20% less than modern.

A third lake of this study, Dune Lake (Chapter 3), sits on an ancient sand sheet and is different hydrologically from the other lakes. Groundwater is a major component of the water budget of Dune Lake and appears to drive long term lake level changes. Present day lake evaporation exceeds precipitation at Dune Lake and a positive flux of groundwater keeps the lake from going dry. Lake-level at Dune Lake has been declining over the course of this study, but prior to this study the lake had been increasing in size and depth since at least the mid-1980's. The ages of drowned trees around the perimeter suggest that lake-levels had not reached such high stands for at least 40-50 years.

Groundwater has been monitored at 5 different wells around the lake since 1995 (when Dune Lake was at a peak high level). Water levels in the wells were declining at a similar rate and magnitude. Groundwater is influenced by many parameters in Alaska. In some drainages, significant water is tied up in glaciers and permafrost is extensive (80% of interior Alaska is underlain by permafrost (Osterkamp *et al.*, 1997)). Melting glaciers and snowpack contribute to stream loading and in turn to groundwater levels. There appears to be some relationship between snow pack, river discharge and changes in groundwater. Permafrost plays a more complex role in influencing hydrology by perching water tables and blocking water flow. When permafrost thaws and melts, the results are often unpredictable (Jorgenson *et al.*, 2001; Osterkamp, 1996; Osterkamp and Romanovsky, 1999). It will take more extensive hydrologic and meteorologic monitoring over a much larger region to definitively identify the factors responsible for changing lake-levels at Dune Lake.

The final two chapters of this dissertation consider recent decadal-scale climate variability and its influence on white spruce growth. The regional climate of Alaska is controlled by several factors. Sudden changes in Alaska climate regimes appear to be promoted by perturbations in large-scale regional atmospheric circulation systems. Physiographic controls in the form of extensive mountain ranges are largely responsible for the dryness in interior Alaska. During the 20th century period of instrument-based climate records, multi-decadal climate regimes have alternated between relatively cool-wet and warm-dry. The hot (post-1970s) regime is the driest sustained period on record. Cool-moist summer regimes reflect a strengthening of southwest maritime flow from the southern Bering Sea into interior Alaska, and hot-dry regimes are characterized by predominance of a high-pressure regime over central Alaska with circulation from the northeast. The 20th century record of climate in interior Alaska shows a cool and wet period in the early to mid part of the century, with increased warming and drying over the past 20-30 years.

The great majority of dendrochronological literature on white spruce (*Picea glauca* (Moench) Voss) in western North America revolves around trees collected at treeline (Ebbesmeyer *et al.*, 1990; Jacoby *et al.*, 1985). However, the greatest biomass of annual white spruce production takes place in lower elevation stands (Ruess *et al.*, 1996) rather than treeline stands. The trees analyzed in Chapters 4 and 5 are from low elevation upland sites throughout interior Alaska and contain an extractable climate signal. Further, they allow a view of the response of the boreal forest to past climate change. White spruce radial growth is consistently negatively correlated with summer temperature throughout the period of record. There has not been a change of sensitivity during the 20th century, as experienced by some treeline sites. Chapter 4 documents that this negative growth effect is caused by temperature-correlated drought stress through analyses of 3 properties of upland white spruce tree-rings: (1) ring-width, (2) ^{13}C discrimination, and (3) maximum latewood density.

Maximum latewood density and ^{13}C discrimination of upland white spruce are most strongly correlated with mean monthly temperatures in the period May through August in the year of ring formation. Chapter 5 depicts a reconstruction (which accounts for serial autoregression and multiple colinearity) based on these two proxies that is highly significant and in excellent agreement with the record of summer temperature during the entire period of instrument record at Fairbanks (1906+). The boreal forest elements in Alaska are thus currently sensitive to moisture stress due to temperature or precipitation fluctuations, and paleovegetation may have been drought-sensitive in the past as discussed in Chapter 5. The 200-year summer temperature reconstruction for interior Alaska reveals four main features. (1) the 19^{th} century in general is reconstructed as being warmer than results of previous studies, which were based on treeline tree ring-width and latewood density. (2) The entire summer temperature record shows rapid changes between 1 to 3 decade-long temperature regimes. (3) Two reconstructed summer temperature regimes (1834-1851 and 1862-1879) are nearly as warm as the post 1970s warm regime. (4) From the 200-year perspective, the early 20^{th} century was a uniquely cool (and relatively moist) period that was the most favorable period in the record for the radial growth of upland white spruce. The summer temperature trends in Alaska both during the periods of instrument data and as reconstructed from tree-ring proxies are often different from mean Northern Hemisphere conditions (Chapter 5).

The overall significance and contributions of this thesis can be summarized as follows:

1) The first quantitative estimates of paleoprecipitation for interior Alaska since the last glaciation were made from data on modem hydrology and past lake level (Chapter 2).

2) Hydrological models for groundwater dominated (Chapter 3) and non-groundwater dominated (Chapter 2) lake systems were developed.
3) A surprising and sustained negative sensitivity in white spruce to summer temperatures was found, which contrasts with previous studies from treeline sites (Chapter 4).
4) A semi-arid Alaska climate, past and present, was confirmed, with implications for boreal ecosystem response to future climate changes (Chapter 2, 4 & 5).
5) Climate in interior Alaska is packaged as sustained regimes over multi-decadal periods. Resource managers and planning boards need to recognize such phenomena in order to properly manage forests (Chapter 5).
6) Future work with multi-proxy tree ring parameters at sites in different environments is necessary to resolve issues of changing tree-growth sensitivity and differences in past climate reconstructions in Alaska.

Based on the results of this dissertation, I might predict that the Alaska landscape would change substantially during the 21st century if anthropogenic warming continues. While we reconstruct temperatures as in the 19th century as warm as modern temperatures, we don't know what might develop with increased sustained warming. In areas where white spruce is already limited by moisture, we might expect to see a transition to an aspen parkland vegetation type, which is now found primarily in the southern extent of the Canadian boreal forest (Hogg and Hurdle, 1995). Lake levels will be harder to predict in the short term, but in interior Alaska a decrease in effective moisture will most likely cause lakes to become shallower. We can expect to see more thaw ponds and lakes formed where permafrost is extensive. Usually such ponds and lakes are of short duration, since eventually the melting is so extensive that the water drains. We need an extensive collection of meteorologic and hydrologic monitoring stations in Alaska to provide baseline data and to document changes.

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[1]Chapter 2

Late Quaternary Paleoclimatic Reconstructions for Interior Alaska based on Paleolake-Level Data and Hydrologic Models

Abstract

Hydrologic models are developed for two lakes in interior Alaska to determine quantitative estimates of precipitation over the past 12,500 years. Lake levels were reconstructed from core transects for these basins, which probably formed prior to the late-Wisconsin. Lake-sediment cores indicate that these lakes were shallow prior to 12,500 yr B.P. and increased in depth with some fluctuation until they reached their modern water levels 4,000-8,000 yr B.P. Evaporation (E), evapotranspiration (ET), and precipitation (P) were adjusted in a water-balance model to determine solutions that would maintain the lakes at reconstructed levels at key times in the past (12,500, 9,000 and 6000 yr B.P.). Similar paleoclimatic solutions can be obtained for both basins for these times. Results indicate that P was 35-75% less than modern at 12,500 yr B.P., 25-45% less than modern at 9,000 yr B.P. and 10-25% less than modern at 6,000 yr B.P. Estimates for E and ET in the past were based on modern studies of vegetation types indicated by fossil pollen assemblages. Although interior Alaska is predominantly forested at the present, pollen analyses indicate tundra vegetation prior to about 12,000 yr B.P. The lakes show differing sensitivities to changing hydrologic parameters; sensitivity depends on the ratio of lake area (AL) to drainage basin (DA) size. This ratio also changed over time as lake level and lake area increased. Smaller AL to DA ratios make a lake more sensitive to ET, if all other factors are constant.

Introduction

General circulation models (GCM's) indicate that high-latitude regions are sensitive to changes in both temperature and P and that significant changes are likely to occur due to the increase in anthropogenically-produced, radiatively-active trace gases (Houghton *et al.*, 1996; Houghton *et al.*, 1990). Signs that changes are already occurring in Alaska include increases in permafrost temperature (Chapman and Walsh, 1993;

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Lachenbruch *et al.*, 1988; Walsh and Chapman, 1990), increased thermokarst activity (Osterkamp, 1996; Osterkamp *et al.*, 1997; Osterkamp and Lachenbruch, 1990) and changes in tree ring characteristics indicating increased stress over the past 20 years or so (Barber *et al.*, 1997; Briffa *et al.*, 1998; Jacoby and D'Arrigo, 1995; Juday *et al.*, 1997). Verification of GCM's and regional climate models can be accomplished by comparing simulations of temperature and P under different boundary conditions with paleoclimate reconstruction based on the geologic record (COHMAP, 1988).

Lakes are a valuable source of diverse paleoclimatic and paleohydrologic proxy data and serve as integrators for whole watersheds. Past lake-level fluctuations can be used for reconstructing and interpreting regional paleohydrologic and paleoclimatic changes on time scales from 10 to 10^6 years (Digerfeldt, 1986; Street-Perott and Harrison, 1985). Closed lakes (lacking outlets) are most sensitive to changes in climate and are generally found in arid to semi-arid regions. Lake-level fluctuations may be recorded by a number of proxies contained in the sediment record. Pollen, diatoms, sediment composition, sedimentary structures and level of the sediment limit have been used to infer lake-level changes. Changes in geochemical conditions associated with lake volume variations can be detected using chemical, mineralogical, isotopic, or biological methods.

Lake-level changes in response to climate change have been well documented in the tropics and temperate latitudes (Digerfeldt, 1988; Digerfeldt *et al.*, 1992; Hostetler and Benson, 1990; Street-Perott and Harrison, 1985) but similar studies are less common in arctic and subarctic environments. In some regions, arctic and subarctic lakes have the potential to provide paleohydrologic records if the climate is sufficiently arid and basin configuration suitable. In order for lakes to exist in semi-arid to arid climates, they must generally be contained within drainage basins that are relatively large compared to the lake area. In these types of systems, runoff is typically the most important component of the water input to the lake. If runoff approaches 100% of total input to the lake, (i.e. extremely large basin compared to lake area) the lakes are known as amplifier lakes, and when runoff approaches 0% of total input, lakes are atmospherically controlled (Street-Perott and Harrison, 1985). Lakes of interior Alaska fall between these two extremes. Here E can exceed P falling onto the lake, while ET from the drainage basin is close to, but less than, P falling on the basin. Thus, runoff is important in maintaining lake water balance, with spring runoff associated with snowmelt having a large impact (Hinzman *et al.*, 1996). The region of Alaska between the Alaska and Brooks ranges was largely unglaciated during the late-Pleistocene and Holocene and lakes in this region have the potential to provide long continuous records of paleohydrologic change.

Northern regions can be classified as subarctic or arctic based on temperature and permafrost (Prowse, 1990). Permafrost is discontinuous in subarctic regions and continuous in arctic regions. Subarctic refers to those regions where mean temperature for the coldest month is below freezing and the mean temperature of the warmest month is above 10°C, but no more than 4 months have a mean temperature exceeding 10°C. The Arctic is classified as the region where mean temperature for the warmest month remains below 10°C and the mean annual temperature is below freezing. The Arctic can be subdivided into low and high regions based on July temperatures. Regions where mean July temperature falls below 5°C are termed High Arctic (HA), while Low Arctic (LA) refers to those regions where mean July temperature falls between 5-10°C. In Alaska, the Low Arctic is generally found north of the Brooks Range where vegetation is in the form of tussock tundra, mosses, lichens and shrubs (dwarf birch, willow and alder). High arctic conditions are not found in Alaska but exist, for example, in the islands of the Northwest Territories, Canada. Here vegetation density is low and there is an increasing proportion of mosses and lichens, which do not transpire, among the tussock tundra and grasses (Prowse, 1990). Precipitation and E (including ET) generally decrease with increasing latitude. Low arctic annual P is usually less than 250 mm while high arctic annual P is typically 150 mm or less. Ranges of E and ET for low arctic conditions are 260-340 (Marsh and Bigras, 1988) and 130-240 mm (Hinzman, 1990; Kane *et al.*, 1990), respectively. In the High Arctic, ranges of E are 84-180 mm (Marsh and Woo, 1979; Woo *et al.*, 1981) and ET ranges from 50-90 mm (Ohmura, 1982a; Ohmura, 1982b).

Interior Alaska (north of the Alaska Range, south of the Brooks Range and east of the Seward Peninsula) is classified as subarctic and the climate is predominantly cold-continental. Summers are short and hot while winters are long and cold. Annual temperature extremes are great (e.g., 38 to -60°C) and permafrost exists over 80% of this zone (Osterkamp *et al.*, 1997). The region is semi-arid, due in part to distance from the ocean and large topographic barriers such as the Alaska and Brooks ranges. Annual P ranges from 120 to 600 mm. The climate of the region encompassing the lakes in this study is classified as semi-arid, cold microthermal, with little or no P surplus, and temperature efficiency normal to cold microthermal (Patric and Black, 1968). Potential annual ET (PET) ranges from 400 to 470 mm. It is highest during the summer, when long hours of daylight, warm air temperatures, and low relative humidity create a high demand for soil moisture and produce a net loss in water resources (Gieck and Kane, 1986). Potential E, based on Thornthwaite's equation (Patric and Black, 1968), is a function of temperature. Actual E (combined E and ET) is often less than potential E and is less than or equal to P (Patric and Black, 1968).

Although interior Alaska is largely forested at present, paleo-vegetation analyses (Ager, 1975; Anderson and Brubaker, 1993; Hu and Brubaker, 1996; Hu *et al.*, 1993) indicate tundra vegetation and suggest arctic conditions prior to 12,000-14,000 yr B.P. (all dates are uncalibrated radiocarbon years). This implies that large changes in factors important to the hydrologic cycle have occurred in this region over this time period.

Synoptic climatic controls that operate at regional and smaller spatial scales across Alaska are poorly understood but could be important for understanding past climates (Mock *et al.*, 1998). The winter climate of Alaska is strongly affected by mid-tropospheric variations of the East Asian trough-ridge system. During winter, a strong upper-level trough is centered to the south and traverses through central Canada (Harman, 1991). Cold air masses and cold-core high-pressure systems predominate north of the trough. Precipitation throughout interior Alaska is normally at its lowest of the annual cycle in winter, with many locations typically receiving less than 25 mm in January. Circulation associated with the Aleutian Low causes higher P in southern Alaska, but most of the P is restricted to the windward side of the coastal mountain ranges (Mock *et al.*, 1998).

Spring and fall climates are short transitions between winter and summer. During summer, interior Alaska is primarily affected by mid-tropospheric variations of ridges and troughs (Mock *et al.*, 1998). Increased solar radiation generally explains the peak of the annual temperature cycle for most locations and the highest temperatures in Alaska are centered in the eastern interior. Warm and dry summers are associated with high pressure ridges extending across central Alaska and the Aleutian Low lying considerably south of Alaska (Streten, 1974). Cold and wet summers occur when low pressure troughs enter central Alaska from the North Pacific or the Arctic Ocean. Also during summer, the Pacific subtropical high is prevalent over mid-latitudes, at times advecting warm air from the south (Mock *et al.*, 1998). Throughout most of interior Alaska, P exhibits a mid-late summer maximum (July and August), as a result of the East Asian trough steering mid-latitude cyclones through the entire region, and also due to a strong Pacific subtropical high advecting some moisture from the northern Pacific Ocean and Bering Sea (Mock *et al.*, 1998).

Study Area

The lakes of this study are Birch (64°18'N, 146°40'W) and Jan (63°34'N, 143°54'W) (Table 2.1). Birch Lake lies within a small east-west trending valley of the Yukon-Tanana Upland. It was probably formed by aggradation of the Tanana River and its tributaries as the discharge and sediment load increased at the onset of the previous interglacial, or possibly during the Wisconsin Glaciation, (Ager, 1975; Blackwell, 1965). AMS dates on seeds from sediment cores indicate that Birch Lake is at least 13-14,000 years old (Bigelow, 1997) and may be as old as 15,000 years (radiocarbon dates on bulk sediment (Ager, 1975)). The lake is enclosed by bedrock and colluvium on all but parts of the western shore, where the dam of outwash sand and gravel was deposited (Ager, 1975; Blackwell, 1965). Today Birch Lake is an open basin with one surface outlet. It is surrounded by boreal forest, has discontinuous permafrost in the drainage basins and probably receives the majority of its water from overland runoff. The surface area of the lake is 3.01 km^2 and the drainage basin area is 37.0 km^2 (Table 2.1). The present outlet is contained by a concrete weir built by the Alaska Department of Fish and Game to control lake level and to prevent fish migration. Birch Lake consists of two basins separated by a sill 7 m below overflow level (Table 2.1). The south basin is deeper with a maximum depth of 14 m, while the north basin has a maximum depth of 12 m.

Jan Lake lies in the Tanana Valley about 220 km southeast of Birch Lake and is a much smaller, closed basin lake with a surface area of 0.143 km^2 and a drainage basin area of 0.643 km^2 (Table 2.1). It is about 12 m in depth and consists of one simple, circular basin. It probably formed in a similar manner to Birch (alluvial damming by the Tanana River) possibly prior to the Wisconsin Glaciation, but more than 12,000 yr B.P. The lake surface lies 60 m above the elevation of the adjacent Tanana River and is surrounded by steeply sloping boreal forest with discontinuous permafrost in its drainage basin. The basin lies on metamorphic bedrock with the alluvial dam along the NE shore.

Table 2.1. Lake characteristics, hydrologic parameters and paleolake levels.

	Birch Lake	Dune Lake
AL (km^2)	3.01	0.143
DA (km^2)	37.00	0.643
Volume (10^6 m^3)	18.91	0.85
Mean depth (m)	6.28	5.95
Median depth (m)	8.34	5.80
Maximum depth (m)	14.0	12.0
Mean ann. air temp. (°C)	-3.44	-5.4
Mean ann. Precip. (mm)	328	250
Mean ann. E (mm)	600	550
Mean ann. ET (mm)	261	183
PET (mm)	449	425
Discharge (10^6 m^3 yr^{-1})	1.7	None
Weather station used	Eielson (1944-1973)	Tok (1954-1984)
Depth at 12.5 kyr. B.P.	3	2
Depth at 9 kyr. B.P.	10	3
Depth at 6 kyr. B.P.	13.4	6

Methods

A modern hydrologic budget for each lake was developed using morphometric and climatic data in a water-balance model. Mean annual water balance of a lake is given by equation (1):

$$\Delta V = AL\,(P - E) + DA\,(P - ET) - D + \Delta G \tag{1}$$

where: ΔV= net change in lake volume

AL = lake area

DA = drainage basin area

P = precipitation

E = evaporation from the lake

ET = evapotranspiration from the drainage basin

D = surface discharge from the lake

ΔG = net flux in ground water

Lake and drainage basin areas were obtained by digitizing outlines from topographic maps. Alaska Department of Fish and Game provided lake bathymetric maps for Birch and Jan lakes. Surveys were conducted with a depth finder equipped with GPS to verify and modify the bathymetric maps. Lake bathymetric contours were generated at 1m intervals and were digitized for area calculations to develop hypsographic relationships. To derive depth-volume and area-volume relations for use in the water balance model, volumes were calculated at each 1m contour by integrating the area-depth curve.

In shallow lakes, sediment infill will cause depth-area-volume relationships to change with time if the configuration of the basin changes. For example, Birch Lake is presently 14 m deep, but the central lake floor was about 18 m below present day level at 12,500 yr B.P. The sediment-core transect and seismic data for Birch Lake (Abbott, 1996) allowed for estimation of lake morphometry for times in the past. Based on this data, depth-area-volume relationships were established for Birch Lake at 12,500 yr B.P. Steady-state water balances for the modern and paleobasins using the same combination of E, ET and P produced similar results (within 5%) for lake area and water depth. Thus it appears that changes in lake morphometry due to infill have little effect on the water depth-area relationship relative to water balance calculations because the shape of the basin changes little. We assumed the same relationship for Jan Lake and therefore have used modern hypsography for past reconstruction.

Lake E and catchment ET are both difficult to measure. Our estimates for E are based on studies from other interior Alaskan sites (Dingman *et al.*, 1980; Hinzman, 1990; Kane and Carlson, 1973; Kane *et al.*, 1979a; Kane *et al.*, 1990; Nakao *et al.*, 1981), from Patric and Black's (1968) calculations using Thornthwaite's equation, from Canadian studies in similar climatic regions (Landals and Gill, 1973; Marsh and Bigras, 1988; Marsh and Woo, 1977; Newberry *et al.*, 1979; Roulet and Woo, 1986) and from empirical calculations (this study) using monthly climate parameter averages (Penman, 1948; Penman, 1956; Ward and Elliot, 1995). The model was used to solve for catchment ET by substituting in the remaining variables assuming steady-state conditions.

For this study we assume the net ground-water flux is minimal or zero. In permafrost regions, infiltration into deep ground-water systems is restricted and runoff assumes a much more important role than ground water in the water balance of most watersheds (Sloan *et al.*, 1975). Permafrost is discontinuous and probably of minor extent in the basins of this study, although its distribution has not been mapped in detail. Even in the absence of permafrost, the relatively impermeable nature of the underlying strata may dicate that ground-water fluxes are relatively small (Street-Perott and Harrison, 1985). If there is ground-water exchange, it flows into the lake in most situations and is negligible compared to the other flows except where spring fed or in caustic areas (Bengtsson and Malm, 1997). For Birch Lake, the majority of the ground-water recharge comes from runoff in the drainage basin (D. Kane, pers. comm.). Thus, the assumption of no net ground-water transfer may be reasonable for these environments although changes in depth of the active layer over time could affect soil water storage.

For closed-basin Jan Lake, discharge is zero. Annual discharge was calculated for the open water season of 1993 at Birch Lake by monitoring the water depth over time in the outlet stream. Data were collected using a pressure logger. Lake discharge at various stream outlet heights was measured with an Epley current meter. The relationship between depth of water in the weir and current discharge was used to estimate total annual outflow from the logged height data.

Meteorological data available for Alaska are somewhat sparse. Data collection sites are located mainly along the coast or the few roads. Most stations around Alaska do not measure parameters necessary to calculate energy budgets. Net solar radiation is rarely collected and cloud cover, when collected, is of limited use because cloud type and height are not recorded, and only percent sky coverage is noted. Precipitation in this region has large spatial variability and so annual P rates measured at one locality, may not necessarily be representative of that region. Precipitation and temperature data were obtained through the National Weather Service from stations closest to the lakes of this study. The nearest weather stations are 50 km (Eielson Air Force Base) and 60 km (Tok) from Birch and Jan lakes, respectively (Table 2.1).

A weather station was established near Birch Lake for most of the open water period of 1993. Temperature, humidity, barometric pressure, solar radiation, wind speed and direction, and P were measured 2 m above the ground. A pressure-temperature (PT) sensor was deployed in Birch Lake during the same period to monitor lake-level changes and temperature at 1 m depth. These were used to study the response of lake level to short term meteorological events in order to estimate ET.

Lake-level reconstructions are based on interpretation of sediment properties with respect to water depth in offshore core transects, with age control determined by AMS ^{14}C dating of pollen and macrofossils (Abbott, 1996; Abbott *et al.*, 1999 (in review)). Seismic profiling also aided in lake-level reconstruction of Birch (Abbott, 1996; Abbott *et al.*, 1999 (in review)). Results indicate that significant changes in lake level occurred over the past 13,000 years (Table 2.1). Both lakes were extremely low from earliest records through 12,500 yr B.P. Lake levels rapidly increased after this time, but with significant fluctuations. At 9,000 yr B.P., Birch Lake was 10 m and Jan Lake 3 m deep. By 6,000 yr B.P., Birch Lake was near present day level while Jan Lake was half as deep as present level (6 m). There is no evidence that past changes in outlet morphology had a significant influence on lake-level reconstruction for Birch Lake. Jan Lake appears to have always been closed.

A water balance model was developed for each lake using a computer spreadsheet program to solve Equation 1. The model incorporated P, E and ET in conjunction with water depth and volume determined from hypsographic relationships. For each lake, we first determined parameters necessary for steady-state at modern lake level. A series of steady-state solutions for different paleo-lake levels was then obtained by adjusting combinations of P, E, and ET. More realistic solutions for P are based on constraining E and ET to values that are likely to reflect conditions at the time determined from pollen data. Calculations were made at a series of time steps (multiple iterations) until steady-state was reached. To illustrate the insight provided by the modeling, we focused on three time periods (12,500, 9,000 and 6,000 yr B.P.). For these solutions, P was set as a percent of modern (125, 100, 75, 50, and 25%) for individual runs of the model.

Results

Evaporation was calculated for Harding Lake, 20 km northwest of Birch Lake by Nakao et al. (1981) using long-term meteorological data from nearby Eielson Air Force Base. Using Penman's formula (Penman, 1948; Penman, 1956), they calculated an annual E of 608 mm from the lake. This indicates that E from interior lakes is greater than annual P, which has been noted by other researchers as well (Marsh and Bigras, 1988). Pan E data from the Agricultural Experiment Station at Fairbanks, Alaska was compiled by Kane et. al. (1979). Using a recommended pan coefficient for semi-arid climates of 0.8 and 0.9 (R. Jaetzold, pers. comm.), we calculated an annual E range of 539-606 mm. Average summer (May-Sept.) temperatures for Eielson, Fairbanks and Delta Junction vary between stations by less than 0.5°C, and since Birch Lake lies between these stations, the use of Fairbanks pan E for Birch Lake is reasonable. Based on these results, an E of 600 mm was used in the models for modern day E for Birch Lake. For Jan Lake, nearby Tok weather data were used. Summer as well as annual temperatures for Tok are 2°C cooler than the station data used for Birch Lake. Based on these temperature data and using Penman (1948 and 1956) and Meyer equations (Ward and Elliot, 1995), we calculated annual E estimates about 5% and 12%, lower, respectively, than that calculated for Birch Lake. This translates to an annual E of between 540-570 mm for Jan Lake and we therefore adopted a value of 550 mm for this site.

Estimates of ET are difficult to obtain directly from field measurements and are typically estimated by difference using equation (1), the approach adopted here. The P, E, and discharge (Birch Lake) values discussed above were used to calculate ET in the model assuming the lakes are now at steady-state (Table 2.1). The estimated ET values are 261 mm and 183 mm for Birch and Jan Lakes, and ET/P ratios are 0.8 and 0.73, respectively. There are different factors that could contribute to the lower value of ET for Jan Lake besides cooler summer temperatures. The basin that drains into Jan Lake is much steeper and has more north facing slope than does the basin that drains into Birch Lake. This would allow for lower transpiration rates and therefore more runoff. Calculations using the Thornthwaite equation also indicate that annual ET/P ratios are close to 1.0 for interior Alaska sites (Patric and Black, 1968).

Another way to estimate runoff (RO) or ET is by making short-term measurements of lake-level rise as a result of individual P events. From equation (1), ET = P - RO (on a volumetric basis). This is a rough approximation of overall rates because ET depends on soil moisture and meteorological conditions at the time of the rainfall event, as well as vegetation type and the duration and amount of rainfall.

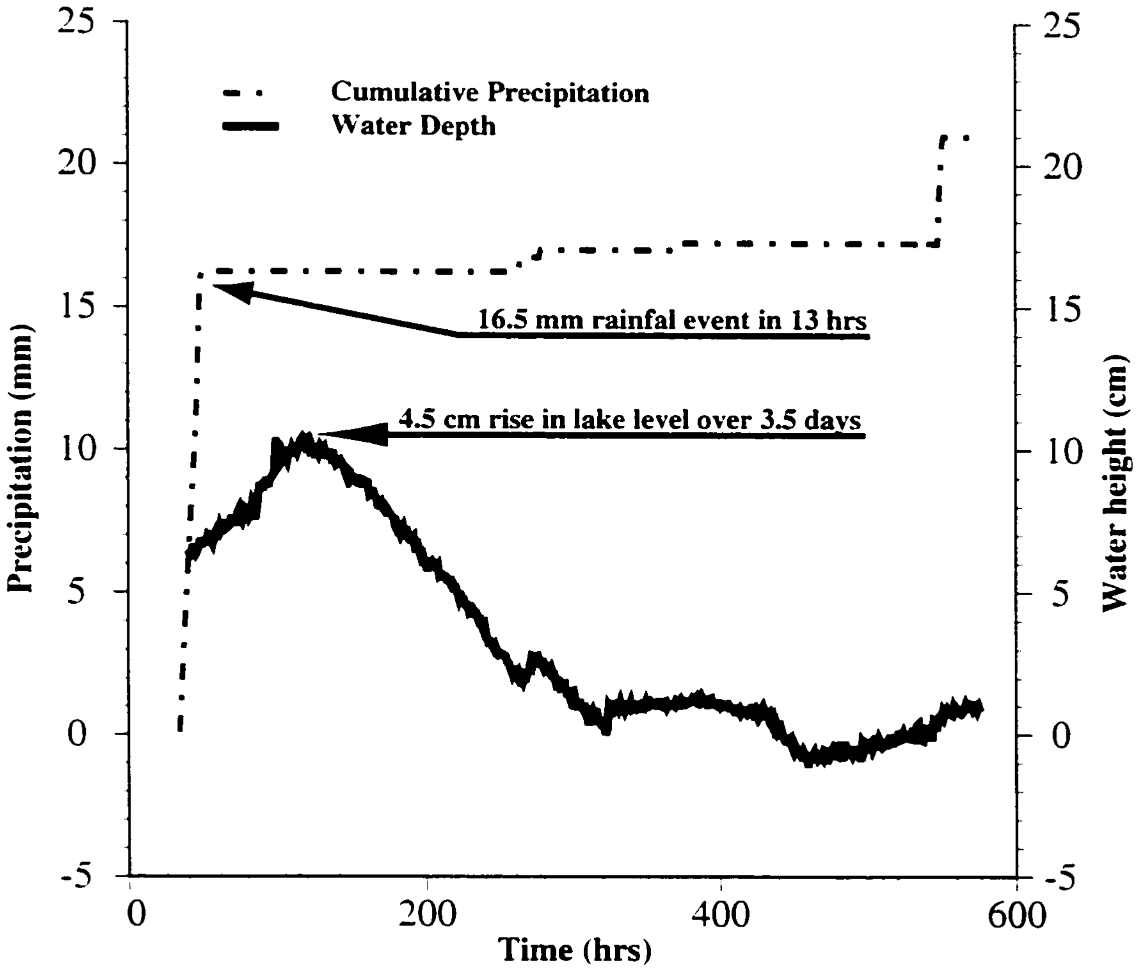

Figure 2.1. Precipitation and lake-level changes at Birch Lake beginning Sept. 17, 1993.

Evapotranspiration will be greater when soils are undersaturated (ET increases with increasing soil moisture up to the point of saturation) and once saturated, runoff will prevail. The response of Birch Lake to a large P event is shown in Figure 2.1. A rapid lake-level rise of 4.5 cm (with a lag over 3.5 days from September 19-22, 1993) is concurrent with a fairly large P event (1.65 cm over 13 hours on September 19, 1993). Evaporation from the lake was assumed to be negligible (rainy, cloudy, cool temperatures) and discharge was zero since the weir was closed. Assuming that ET accounted for the difference between basin rainfall and runoff, and accounting for rainfall directly on the lake, the ET/P ratio for this event was 0.85. While this is only a single event and some soil storage probably occurred, such high ET rates are consistent with previous estimates (Rouse, 1990).

Sensitivity tests were run for each lake to determine the response of lake level to changes in P, E and ET using the water balance model. While E and ET were held constant, P was changed as a percentage relative to present day P (Figure 2.2a). Since Birch is not a closed basin, any P greater than present does not significantly change lake level; excess water is lost as surface discharge. A precipitation decrease of approximately 13% of modern would turn Birch Lake into a closed basin. Lake level would drop by 10 m in 80 years with a drop in P of 20% less than present. Jan Lake, a closed basin, responds to both positive and negative changes in P. With increases of 10% and 20%, lake level rises 1.5 m and 3 m, respectively, and the new equilibrium is obtained in less than 20 years. With 10% and 20% decreases in P, lake level reaches a steady state in about 60 years at 3.5 m and 10 m, respectively, below present level.

While it may be more realistic to change E and ET together, we conducted tests to highlight the sensitivity of lake level to each factor independently. Precipitation and ET were held constant while E was manipulated (Figure 2.2b). The response of both lakes was not dramatic. Birch Lake level did not change at all, while Jan changed by less than ±2 m with a ± 20% change in E. A ±20% change in E corresponds with a ±4-5°C change in temperature using the empirical relationships for E discussed above. Holding P and E constant and varying ET by the same percentages resulted in greater change than in manipulating E (Figure 2.2c). Birch Lake did not respond to a ± 10% or -20% change in ET, but dropped by over 2 m with a 20% increase in ET. Jan Lake level rose 1 and 2 m, respectively, with a decrease in ET of 10% and 20%. It fell 1.5 and 5.5 m, respectively,

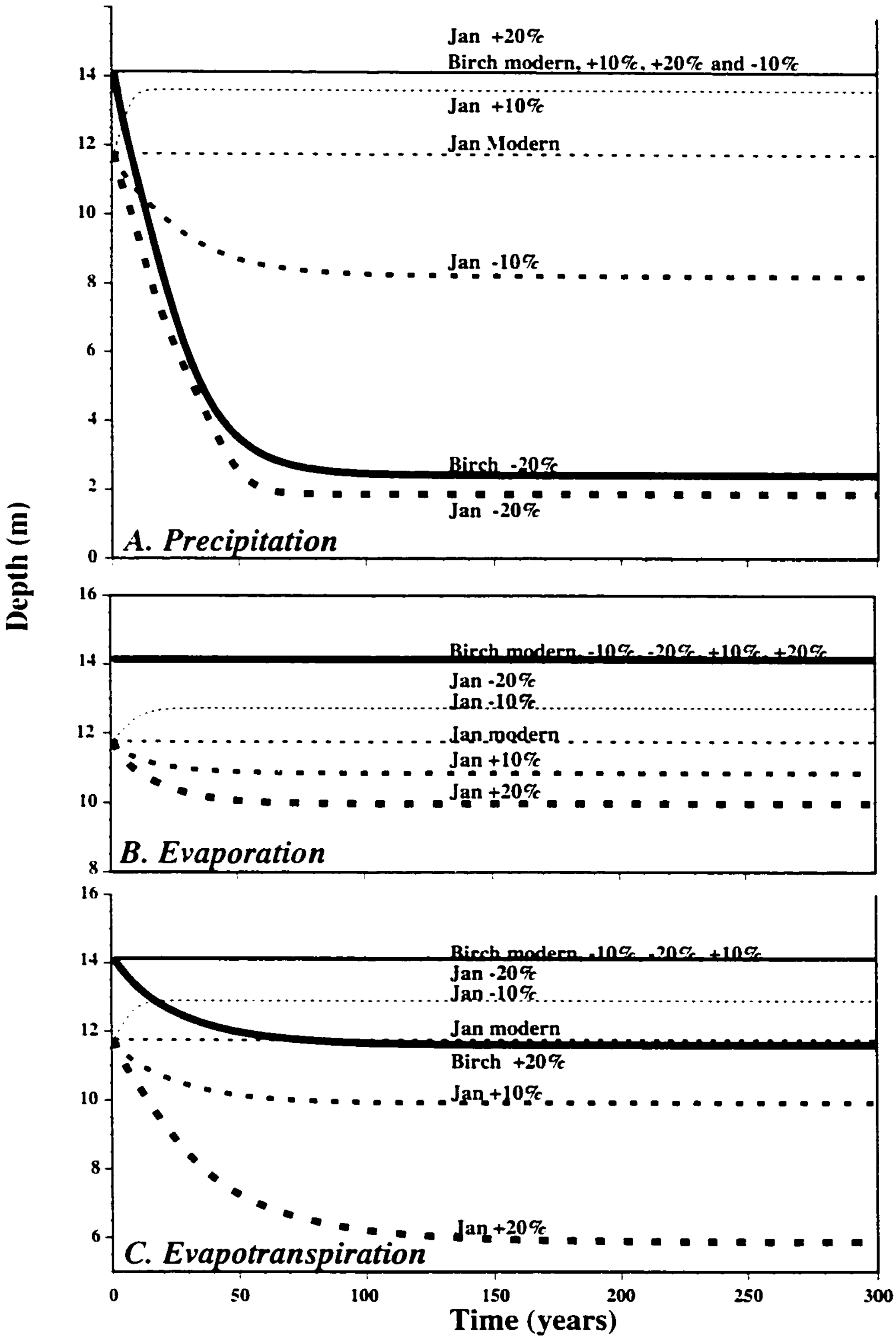

Figure 2.2. Model sensitivity experiment for Birch and Jan Lakes.

with a 10% and 20% increase in ET. The new steady-state was obtained within about 100 years. This exercise shows that these lakes are more responsive to ET than to E. While it may be reasonable to assume that E and ET would change concurrently, there may be a decoupling as vegetation response lags climate change.

Precipitation is the most well constrained parameter for interior Alaska and was held constant at modern values in the following exercise. Steady-state solutions to balance each lake at its modern level using P from Table 2.1 and adjusting E and ET are shown in Figure 2.3. Because Birch Lake is not a closed basin, the measured outflow was held constant. Theoretically, present day lake level could balance at any point along the E vs. ET line. Our estimates of E and ET for Birch and Jan (Figure 2.3) fall within the field of estimated values for E and ET for subarctic sites (Gurney and Hall, 1983; Patric and Black, 1968). Evapotranspiration is somewhat higher for Birch Lake and could be overestimated if discharge is underestimated or if net ground-water discharge is significant.

Using the model and lake-level data for each lake, the paleohydrology of both lakes was reconstructed. The parameters P, E, and ET were manipulated to obtain a set of steady-state solutions for the reconstructed lake level for each time in the past. In the graphical representation of these solutions, P was changed as a percent of modern (Figure 2.4). Evaporation and ET were calculated to determine values necessary to maintain the lake at the given level for each P value. Thus each line represents a contour of constant P (as some % of modern) to balance the lake at the appropriate level for a range of E and ET values. Realistic solutions for P can be determined by constraining E and ET to values based on environmental reconstructions from pollen data for the appropriate time period. The symbols in Figure 2.4 indicate values of modern E and ET estimated by this study for the individual lake sites (M), and average values determined for present day low and high arctic (LA and HA, respectively) sites from previous studies (Hinzman, 1990; Kane *et al.*, 1990; Marsh and Bigras, 1988; Marsh and Woo, 1979; Ohmura, 1982a; Ohmura, 1982b; Woo *et al.*, 1981).

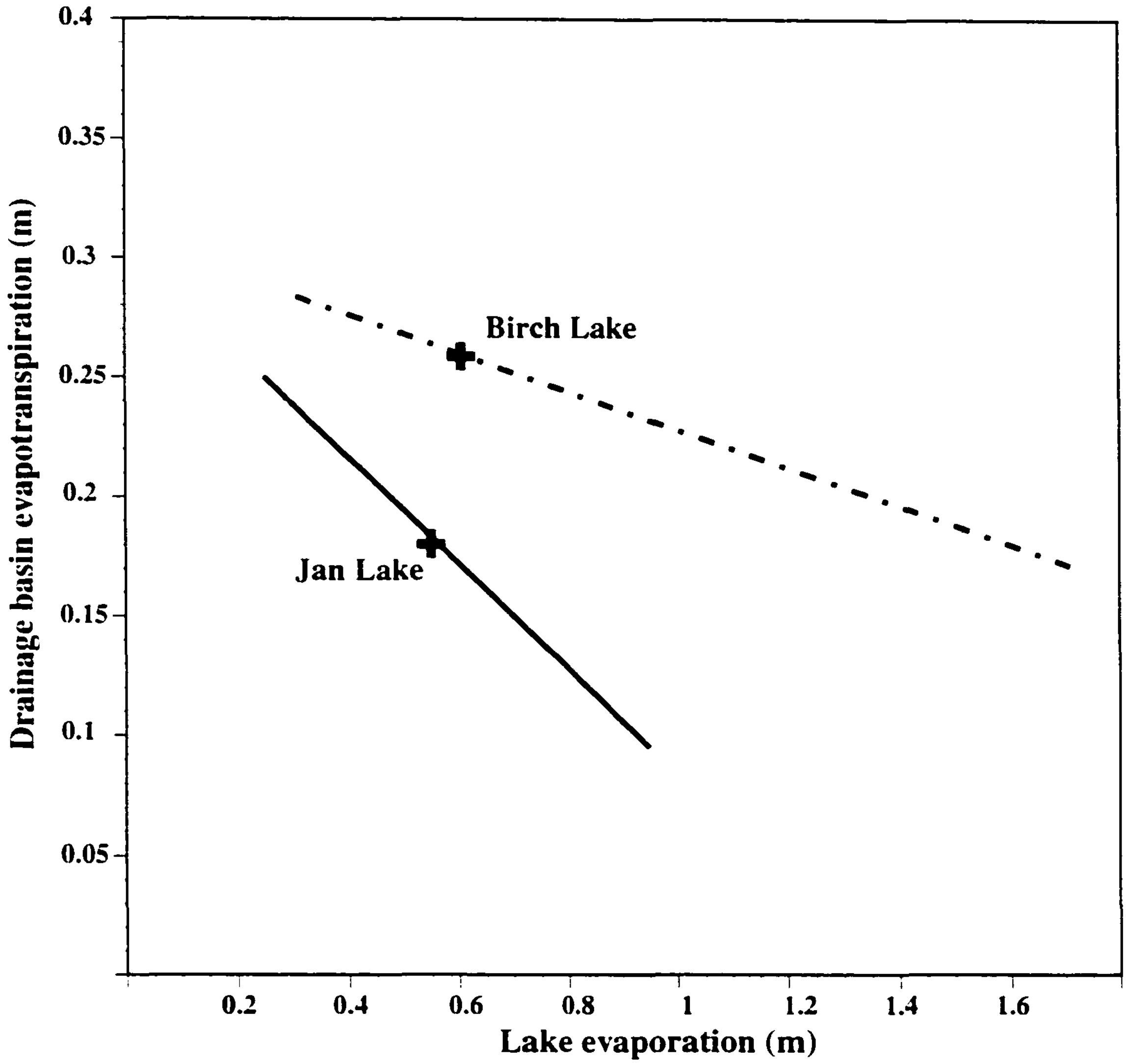

Figure 2.3.A line for each lake indicates modern pecipitation with a range of solutions of evaporation and evapotranspiration to balance the lake at present-day levels.

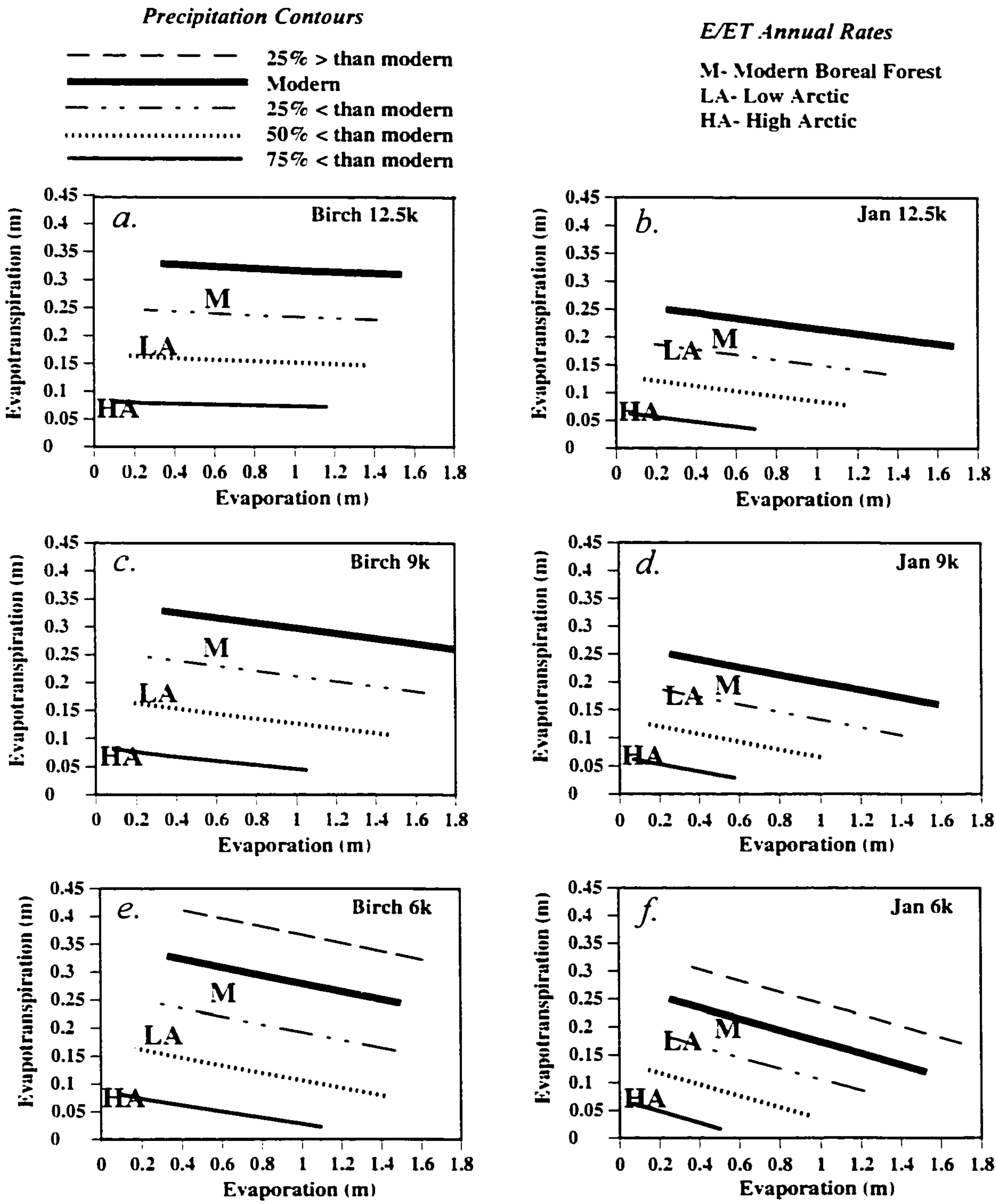

Figure 2.4. Model runs of each lake at three different time slices.

Discussion

At 12,500 yr B.P., Birch and Jan lake levels were low (Table 2.1). The P contours in Figures 2.4a and 2.4b indicate that the range of possible ET values is substantially less than the range of possible E values. This implies that lake level is much more sensitive to ET than E and is likely due to the small lake area-to-drainage basin (AL/DB) ratio. This is particularly true for Birch Lake, which has the lower AL/DB ratio (0.08). For both of the lakes at this time, P would have to have been about 20% less than present to hold the lakes at these low stands under modern day E and ET values. However, at 12,500 yr B.P., E and ET values were probably significantly lower than present, as temperatures were cooler and trees were absent from the landscape. Pollen studies indicate that vegetation at this time was herb tundra with some birch (probably shrubby species, *Betula glandulosa* or *B. nana*; (Ager, 1975; Bigelow, 1997)). Thus E and ET values would have been more similar to LA, or even HA values. Using these values (i.e. the values of the P contours within the LA and HA fields in Figure 2.4a and 2.4b), P would have been much lower for both lakes, with model results 35-75% less than modern. It is highly unlikely that changes in E alone could account for the lower lake levels. Increased winds and decreased humidity could increase E rates, but cooler temperatures would probably offset this. Even if E were twice as large as modern values, a large decrease in P would be necessary to account for the low lake levels. Such low values of P are consistent with inferences based on pollen data from Birch and Harding lakes (Ager, 1975; Bigelow, 1997; Nakao, 1980).

Analysis of lake-sediment cores suggest that interior Alaska lakes were at low to intermediate levels at 9,000 yr B.P. (Table 2.1). Based on pollen data, the vegetation at this time was predominantly *Betula* and *Salix* with some *Populus*, which suggests appropriate ET estimates may have been similar to the present day Low Arctic. With some trees on the landscape and warmer conditions, active layer depths, soil diffusivities, and ET rates would have been greater than at 12,500 yr B.P. Using low Arctic ET and E rates, P estimates (to balance the lakes at their respective levels) range from 25-45% less than modern (Figure 2.4c and 2.4d). By 9000 yr B.P., orbital variations may have resulted in increased seasonality at 65°N, which would have produced warmer-than-present summers but cooler winters (Bartlein *et al.*, 1991; Berger, 1978). This may have resulted in E rates similar to, or greater than, present day subarctic values, but the paleoclimatic data for interior Alaska are somewhat ambiguous on this point. Our sensitivity tests and model results indicate that if summer temperatures were 5°C warmer than present, increased E alone could not account for the lower-than-present lake levels. An E value based on 5°C warmer summer temperature combined with low arctic ET values (from the literature) results in a P estimate of 20-40% less than modern. We conclude that P at 9000 yr B.P. must have been greater than P at 12,500 yr B.P., but lower than present and summer temperatures could have been warmer.

By about 6,000 yr B.P., the sediment record indicates that Birch Lake was near modern day lake level and that Jan Lake was about half its present level (Table 2.1). Birch Lake was assumed to be full but closed (i.e. no discharge) to establish the limits of E, ET and P for which a lake-level drop would not occur. While this delimits a lower boundary for P in the model, values would be greater than model results if there were discharge. The lines of constant P (Figures 2.4e and 2.4f) are more steep than at 12,500 yr B.P., indicating that lake level is not as sensitive to ET at 6,000 yr B.P. as at 12,500 yr B.P. This is due to the larger AL/DB ratio that results from higher lake levels. Pollen records suggest vegetation was similar to the modern vegetation (boreal forest), although *Picea* abundance was lower than present (Bigelow, 1997). This implies that ET would have been higher at this time than for the 12,500 and 9,000 yr B.P. reconstructions, and more similar to modern values. Both lakes balance at their reconstructed 6,000 yr B.P. levels, with 10-25% less than modern P using modern values for E and ET. The increase in lake levels since 9,000 yr B.P., coupled with the probable increase in ET due to the arrival of *Picea*, strongly suggest a real increase in P relative to the previous periods, although P appears to have been still slightly less than present. As with 12,500 and 9,000 yr B.P., a mutually compatible paleohydrologic solution exists for both lakes. A regional P gradient different from today's is not required, even though each lake appears to have a different lake-level status.

Our results show major changes in P and effective moisture (potential E plus ET minus precipitation) between the Late Pleistocene and the mid-Holocene. Based on recorded climate data (1909-1998), interior Alaska on average receives approximately 37% of the annual P as snow and 63% as rain. If this ratio was similar during the Late Pleistocene and the early Holocene, results imply drastically reduced precipitation which would have strongly influenced vegetation and ecosystem processes. Spring melt with accompanying runoff is probably the single most important source of water to arctic lakes. A much smaller proportion of runoff is associated with summer rainfall, except in the case of severe storms (Gieck and Kane, 1986; Kane *et al.*, 1979b). Therefore, changes in the proportion of rain vs. snow could affect lake levels. A long-term increase in the proportion of winter vs. summer P, with consequent increased spring runoff, could cause lake levels to rise. Such changes in the seasonality of P could also lead to conflicting paleoclimatic inferences derived from different data sources. For example, in the scenario of increased winter P and decreased summer P, lake levels may be expected to rise due to the increase in annual runoff, and would imply wetter conditions. However, vegetation may respond to drier summers and suggest more arid conditions.

Freeze-up of interior Alaskan lakes usually occurs in mid to late October, and spring break-up occurs in mid to late May; the ice-free period is about 5 months. Climate changes may result in changes in the length of this ice-free period. Our model results suggest that changes in E resulting from changes in the length of the ice-free period would have a relatively minor effect on lake levels (Figures 2.2b, 2.4). Changes in ET resulting from changes in climate would probably have a much greater impact on lake water budget through impacts on runoff. Changes in vegetation due to climate change would also affect ET rates although there would likely be a time lag.

Conclusions

Modern hydrologic models for lake basins allow for the reconstruction of regional past effective moisture when combined with paleolake-level data. While better long-term hydrologic and meteorologic data for calculating lake water budgets would help refine the models, the results presented here provide insight into past climatic conditions. Estimates of modern hydrologic balance indicate that annual rates of E and ET for two interior Alaskan lakes fall within a range of 420-630 mm and 180-260 mm. The difference between these estimates may be due to differing wind regimes, air and soil temperatures, vegetation, lake and drainage basin size, topography, aspect and patchiness of permafrost.

At 12,500 yr B.P., steady-state solutions for a set of likely E (lake) and ET (catchment) values indicate that P was 35-75% less than modern values. Pollen data indicate herb tundra and imply much lower rates of ET. Evaporation rates were also probably lower than the present. With significant reductions in both E and ET, the conclusion of much lower P is rational. The 9,000 yr B.P. steady-state solutions for lower-than-present lake levels suggest that P was 25-45% less than modern. Evaporation could have been higher than modern due to greater summer insolation, but this factor alone cannot account for the lower lake levels. Evapotranspiration rates would probably not have been as high as modern since trees were less abundant. Decreased continentality as a result of the inundation of the Bering Land Bridge may have increased nearby sources of moisture and probably would have increased P relative to 12,500 yr B.P. Vegetation similar to modern was established by 6,000 yr B.P. Model solutions suggest P 10-25% less than modern which can account for intermediate to high lake levels recorded at this time.

A different lake-level status (i.e. high, moderate, low) can occur in different basins under the same climatic conditions depending on the ratio of lake area to drainage basin as well as other factors as discussed above. The results show that a consistent paleoclimatic solution can be obtained for both lakes at each time slice. A regional climatic gradient different from the present may have existed, but is not required by the results. We can argue conclusively that P must have been less than present in interior Alaska during the Late Pleistocene and early Holocene. Very large changes in P are indicated across the glacial-interglacial transition. The hydrologic model presented here allows for the evaluation of changes in factors such as E and ET, and provides quantitative estimates of P. This is the first work to present quantitative estimates of these changes over millennial time scales and regional spatial scales in Alaska. Results should contribute toward the evaluation of paleoclimatic model simulations and to an understanding of ecosystem response to changing climate in Alaska.

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Chapter 3

Hydrologic Controls on Dune Lake, Alaska: Implications for Recent and Holocene Lake-level Changes.

Introduction

Geological setting

Dune Lake (64°25.5'N, 149°53.83'W) lies at the southern edge of a large dune field north of the Alaska Range in interior Alaska (Figure 3.1). This dune field is part of an extensive system of dunes and sand sheets that covers 4800 sq. miles (Collins, 1985). The source of the sand is glacially-derived alluvium transported northwards towards the Tanana by rivers draining the Alaska Range. It appears that the dune field was formed during the last glacial period (Lea and Waythomas, 1990). As climate became cooler and drier following the last interglacial (stage 5) and the vegetation diminished, winds began to move the sand southwest across the alluvial deposit. The dune field probably stabilized during the early Holocene. Lacustrine indicators in Dune Lake sediment suggest significant dune migration had ceased around 11,000 yrs BP (calibrated yrs)(Bigelow, 1997).

There are two main tracts of dunes, the Northwestern and the Southeastern Tracts. The Southeastern Tract has 3 separate fields, an eastern, central and southwestern area. Dune Lake lies in the eastern area of the Southeastern Tract, near its southern end, and is surrounded by a sand sheet about 7 mi. long by 5 mi. wide. The eastern area of the Southeastern Tract ranges in elevation from 300-700 ft a.s.l. The highest parts of this dune field are the bluffs on the northeast side of the lake. The dunes around the lake are unique to the whole dune field and are rosette shaped, probably due to the local wind regime (Collins, 1985).

Climate

Climate in this region is similar to that of the Fairbanks station (this chapter), although summer rainfall is higher in the Dune Lake region. Winds are variable and include strong dry northeasterlies blowing in the winter, southwesterly tracking summer storms, and occasional very strong localized winds blowing from the southeast through the Nenana River canyon (Collins, 1985).

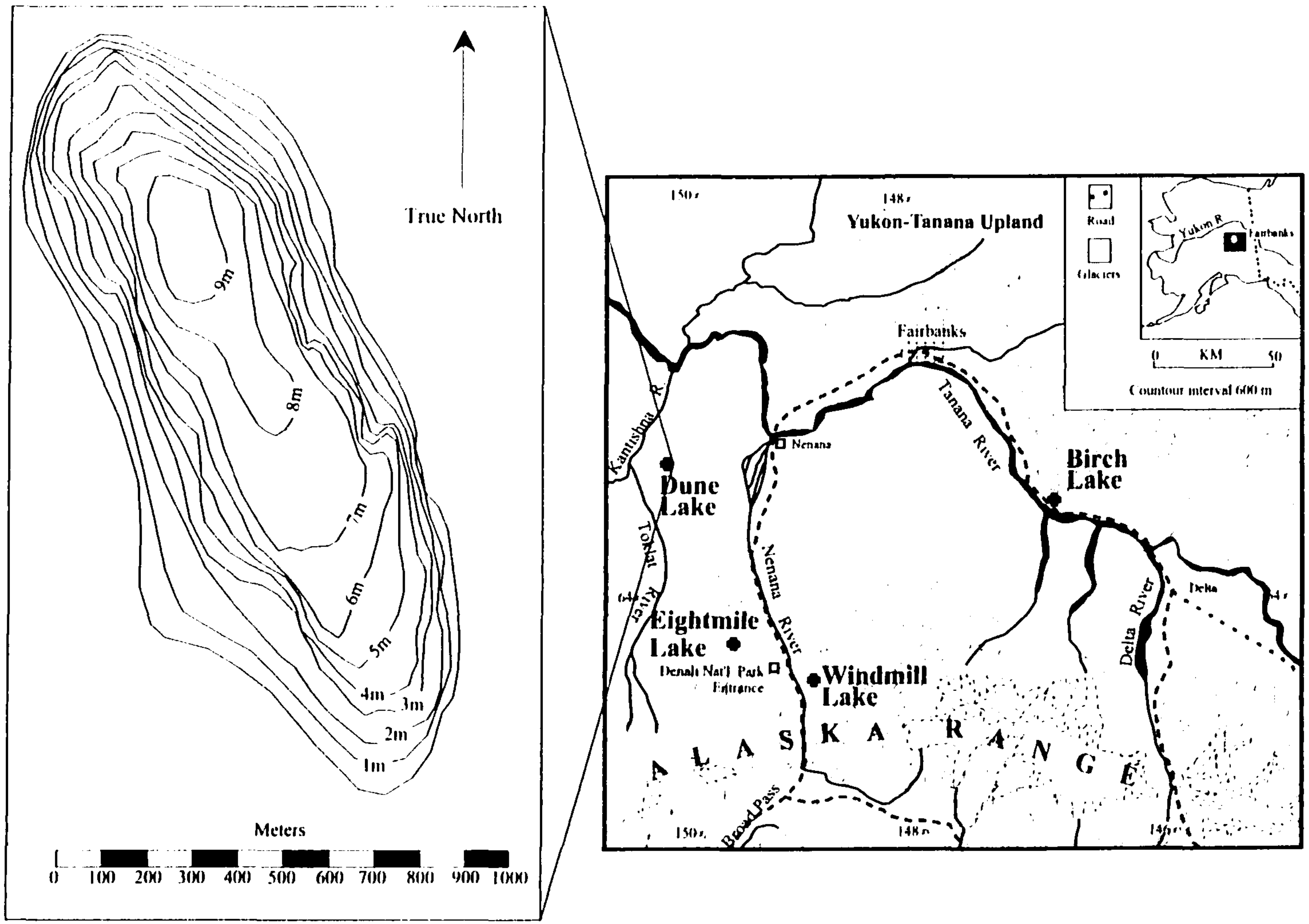

Figure 3.1. Location and bathymetric map of Dune Lake

Vegetation and Fire history

The modern vegetation around the lake is a mixture of open woodland/scrub and mixed closed boreal forest. An extensive fire in the region in 1981 caused about 720,000 acres to burn south of the lake where scrub (*Salix* spp., *Populus tremuloides* and *Betula papyrifera*) grows today. The well-drained beach ridges, south facing dune slopes and dune crest are covered by open woodland primarily *of P. tremuloides, B. papyrifera, P. glauca, Salix spp.* and *Shepherdia* bushes along with lowbush shrubs and grasses. The closed boreal forest occupies the less well-drained areas and the north-facing dunes. Vegetation here is dominated by *B. papyrifera, P. glauca* and occasional *P. mariana* with a shrub layer of *Salix* and *Rosa*. Poorly drained areas are covered by *P. mariana* with a moss layer. (For a more extensive coverage on vegetation, see Bigelow 1997).

Lake-level changes

In 1975, the Alaska Department of Fish and Game did a lake survey of Dune Lake and measured the maximum depth to be 6m, although they might have missed the deepest depression in the lake. There is a recent documented history of lake-level rise since at least the mid-1980's (personal communication) culminating in a maximum depth of about 9m in 1995. This recent lake-level rise resulted in drowned trees and subsequently submerged stumps all around the lake. At present there are about 8 cabins around the lake, which are used for recreation. Since 1995 there has been a decline in lake-level to the present (October 2001).

A series of raised beach ridges of unknown age surround the lake (the highest at ~10m above lake-level). Radiocarbon dates from ancient lakeshores lying above present day lake elevations of another lake in that region give dates of about 20,000 BP (Collins, 1985), but the relationship of this high stand to Dune Lake high stands is unknown.

Hypothesis

Dune Lake appears to be part of a regional groundwater system. Rivers drain the Alaska Range to the north and occupy the low lands surrounding the dune fields. Lakes occupy depressions in the sand sheets and are outcroppings of the groundwater. The groundwater must be fairly close to the surface in order for these rivers and lakes to exist. Any changes in climate affecting groundwater would affect lake-level.

I hypothesize that recent changes in the lake-level of Dune Lake are controlled primarily by alterations in the hydrologic controls of its water balance due to changes in groundwater flux as a result of local and regional weather. It is possible that there are secondary effects due to changes in evapotranspiration (ET) as a result of the large fire of 1981 that might have also affected the hydrologic budget and hence lake-level. Meteorological and hydrologic data were collected from 1995-2000 to test these hypotheses.

Methods

Meteorological Data:

In July 1995, a meteorological station was set up at Dune Lake on a flat open surface about 50 m from the NW shore of the lake and 2 m above the ground. Sensors measured temperature ($\pm$0.005°C), relative humidity, insolation ($\pm$0.01 w/m^2), barometric pressure ($\pm$0.02 cm), wind speed ($\pm$1.0 m/s) and direction and warm season rainfall ($\pm$0.5mm). Data were collected through December 2000 with some breaks in data collection due to equipment malfunction. The most complete data sets included the years 1995-1998. In June 1998 the wind sensor and rain gauge were accidentally disconnected and were reinstalled the following year. The data in 1999 are sporadic and the MET station ceased functioning on January 1, 2000.

Hydrologic Data:

In July 1995, a pressure and temperature sensor was deployed in the lake to monitor lake-level and temperature in conjunction with the meteorological data. The accuracy of the pressure sensor is $\pm$0.02 cm and for temperature is $\pm$0.005°C. The pressure data were used to measure lake-level over time and the temperature was used as an estimate of surface water temperature for calculating lake evaporation. Lake-level is determined from the lake pressure and temperature sensor by making corrections for water density at *in situ* temperatures and for changes in atmospheric pressure using the hourly barometric pressure collected by the meteorological station.

Galvanized steel sand points (1.5 inch inside diameter) were hand-driven approximately 6 feet into the ground at five locations around the lake in 1995 and 1996 (Table 3.1). The relative position of the well head, lake surface, and ground-water level for each site were surveyed once a year using a level and stadia rod (±0.02 cm precision) in 1996-1998 and in 2000 and 2001. Temporary benchmarks (nails driven into trees or stumps) were included in each well survey. Identical measurements between the well heads and temporary benchmarks for all surveys indicate that the well casings were not displaced due to frost heaving. Absolute elevations for each well were not determined, as permanent benchmarks were unavailable, but all surveys were calibrated to one well casing from 1996. Well water levels (groundwater with respect to well head) were measured at least once a year except for 1998, when they were measured three times.

Table 3.1. Location of groundwater wells installed around Dune Lake.

Well	**Latitude**	**Longitude**	**Distance from lake (m) (1997)**
Rick's	64.42229°	-149.8927°	7.75
Southeast	64.4145°	-149.887°	12.96
Southwest Pit	64.4131°	-149.8923°	22.2
West	64.4201°	-149.9033°	17.95
Miller's	64.42856°	-149.8923°	17.3

The water level in the south pit and Miller's wells were monitored continuously from July 1 to October 7, 1997 with an Omnidata water level recorder and data logger and a 0-15 psi pressure transducer. The well data were used to estimate groundwater flux and direction. The yearly surveys provided snapshots in time while the continuous monitoring of South and Miller's wells provided a detailed look at seasonal groundwater variability.

Measurements of water temperature, pH, conductivity, turbidity and dissolved oxygen were made with a Horiba water quality instrument at least once a year between 1995 and 2000. Alkalinity and hardness were measured twice in 1995 and once in 1996. Measurements were taken at 1 m increments in the deepest part of the lake and also on well water samples. A water chemistry survey was also conducted by the Alaska Department of Fish and Game in 1975.

In 1995, water samples were collected from several depths in the lake, from the wells, and from rivers draining the Alaska Range near Dune Lake and $\delta^{18}O$ was measured. In 1997, water was again collected from several depths in the lake and from the wells, and analyzed for $\delta^{18}O$ and the ratio of D/H (deuterium/hydrogen).

A bathymetric survey was done for Dune Lake in 1995 using a recording depth finder and GPS. Hypsographic relationships were developed for Dune Lake by using depth contours at 1m intervals. An area:depth curve was established and the curve integrated for the depth:volume relationship.

Results:

A continuous lake-level curve was obtained from July 1995-October 1998 (Figure 3.2), which shows an overall trend of declining lake-level over this period. Lake-level did rise briefly in 1998, but subsequently lake-level continued to drop. Other data collected in 1999 and 2000 show persistence in this trend. Lake-level in October 2001 is at the lowest level since 1995.

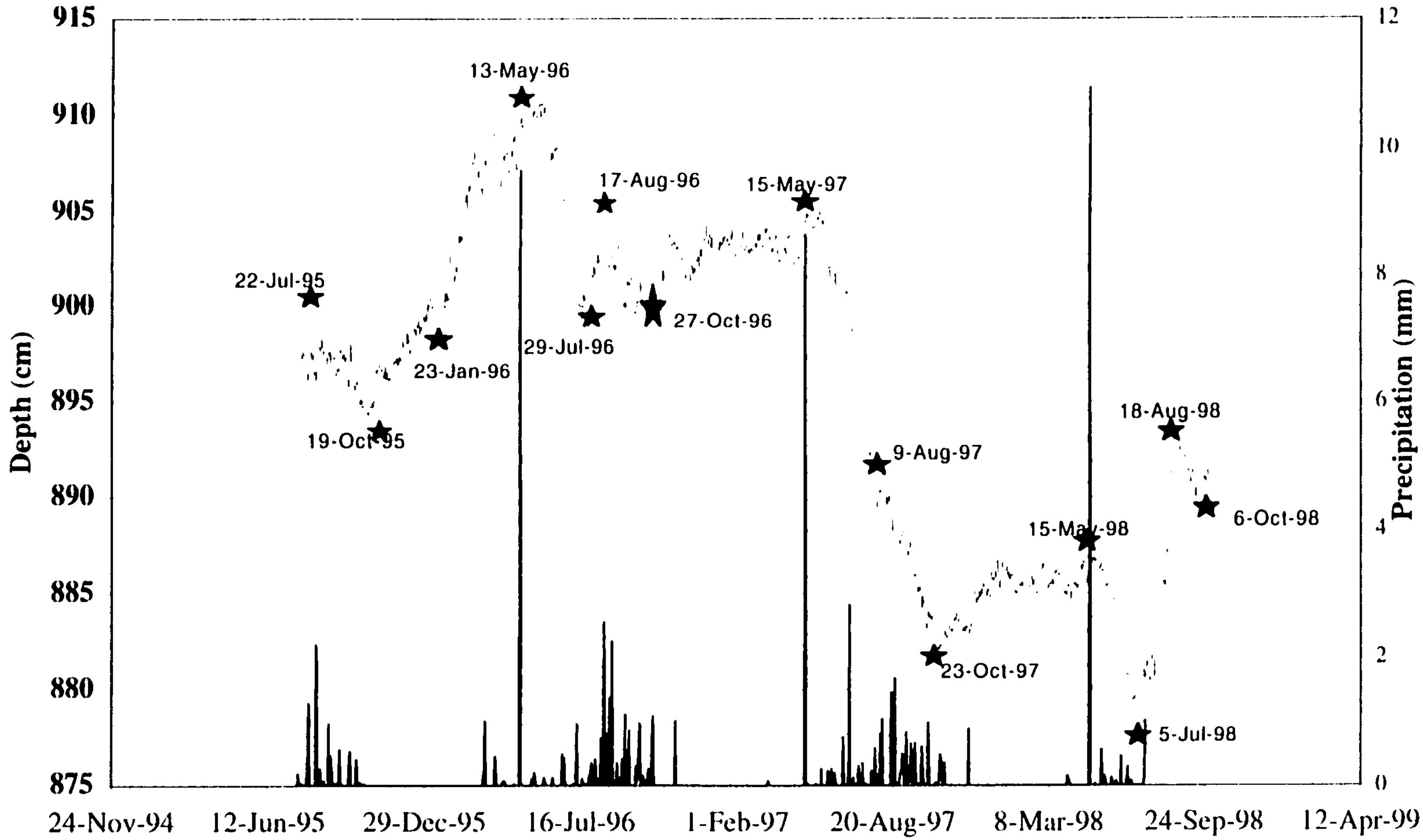

Figure 3.2. Dune Lake level from July 1995 - October 1998, and summer precipitation during this period. Star symbol denotes position related to date.

In an effort to determine the factors responsible for changing lake-levels, lake water balance was calculated on a year-to-year basis for 1995-1998. The water balance Equation is as follows:

$$\Delta V = LA\,(P - E) + R - D + \Delta G \qquad (1)$$

where: ΔV= net change in lake volume

LA = lake area

P = precipitation

E = evaporation from the lake

R = runoff into lake

D = surface discharge from the lake

ΔG = net flux in ground-water

Lake area and seasonal precipitation were directly measured but lake evaporation and groundwater flow were estimated as discussed below. As Dune Lake is a closed basin and there are no visible surface inputs (drainage channels) or outputs (streams or creeks), R and D both = 0. The sand surrounding the lake is fine-grained and water percolates through rather than flowing on top of the surface as runoff.

Thus the Equation simplifies to:

$$\Delta V = LA\,(P - E) + \Delta G \qquad (2)$$

We know that evaporation is important to the hydrologic system and is taking place as the $\delta^{18}O$ signal of the lake water is enriched (-14 ‰) relative to the groundwater coming into the lake (-19 to –21 ‰) and precipitation falling onto the lake. As I will show, the groundwater enters the lake from the south and has a similar isotopic signature as the rivers draining the north side of the Alaska Range in the Dune Lake region.

Miller's well at the north end of the lake has the same isotopic signal as the lake (-13 to –14 ‰) adding further evidence that the flow of water is to the north (Table 3.2).

Table 3.2. Well, river and Dune Lake δ O^{18} values (‰)

Sample water	1995	1997	1999	2000
Dune surface	-13.85		-13.3	-13.2
Dune profile				
2m		-13.34		
3m	-13.95			
4m		-13.47		
7m	-14.07			
8m		-13.78		
8.5m lake	-14.44			
Wells				
Miller's Well	-13.65	-14		
Rick'a well	-14.02	-14.68		
SE well	-21.23	-20.66		
SW Pit	-19.34	-19.44		
West Well	-20.17	-18.85		
Rivers				
Nenana River	-21.22			
Toklat River	-21.13			
Teklanika River	-21.02			
Savage Creek	-20.67			
Sanctuary River	-20.99			

Meteorological data from 1995-1998 were used to calculate lake evaporation using the Penman-Van Bavel method (Van Bavel, 1966) and the Bowen Ratio Energy Balance (Bowen, 1926) using programs written by John Fox (University of Alaska Fairbanks).

The Penman method (Penman 1948, 1956) is a combination of the energy balance Equation and vapor flux approach to calculating evaporation. It is commonly used for time periods ranging from daily to monthly. A refinement to the Penman method was implemented by van Bavel (1966), making it possible to calculate evaporation on an hourly basis and eliminating the dependency on a fixed, empirical wind function (Equation 3).

$$Qe = -(Qnet + Qw) + Qh = 0 \qquad (3)$$

Where:

Qe = latent heat flux (W/m^2)

$Qnet$ = net radiation (W/m^2)

Qw = change in water heat storage (W/m^2)

Qh = Sensible heat flux (W/m^2)

The Penman-van Bavel method utilizes wind speed, air temperature, relative humidity and solar radiation data collected hourly at Dune Lake, while eliminating the need for temperature and vapor pressure measurements at more than one height above the ground. However, net radiation and surface roughness must still be estimated for the calculation. Net radiation required estimating lake albedo and lake longwave emissivity from the iiterature. Lake albedo was estimated as a function of the sun angle from the time and latitude at Dune Lake and the refractive index for water. Longwave radiation from the lake was calculated by using measured lake surface temperature at Dune Lake and the Stephan-Boltzmann law. Incoming longwave radiation was estimated from air temperature and sky-cover class. Sky-cover class was estimated by using data from the Fairbanks climate record. Monthly regressions of global radiation and sky-cover class from Fairbanks were used to calculate sky-cover class from global radiation measured at Dune Lake.

In spite of improvements over the original Penman Equation, there are still some inherent limitations and assumptions in the calculation. The H term was defined by van Bavel as all energy fluxes except for latent and sensible heat flux and includes radiative flux, soil heat flux and heat storage changes in water. This implies that any ground heat flux under the lake or any change in energy stored in the lake should be included in the H term. For Dune Lake we assumed no net change in energy storage during the time increment of calculation (1 hour), which is the equivalent of saying that net radiation is the only component of H. Second, the use of a single wind speed measurement was facilitated by assuming a logarithmic wind profile. The latter is valid for neutral atmospheric stability conditions. Additional assumptions are of negligible vertical divergence of the convected fluid (air) between the surface and the levels of measurement and of the equality of turbulent transfer coefficients for water vapor and sensible heat.

The advantage of the combination method is the algebraic formulation that dampens or buffers errors in the turbulent transfer coefficient or vapor pressure deficit of air caused by measurement errors in wind speed or vapor pressure. Van Bavel showed that a 10% error in the turbulent transfer coefficient or saturation vapor pressure deficit resulted in only a 0.5-1.0 % error in evaporation. Cloudy days were not included by van Bavel when comparing hourly totals with estimates for daily evaporation using average daily data. Also, he found that hourly results were not accurate at night, possibly due to non-neutral stability conditions. In view of the uncertainties of nighttime application of this Equation, when hourly net radiation was negative, the evaporation rate was set to zero for calculations. The rate of evaporation or condensation occurring under conditions of negative net radiation are expected to be small and tend to balance out for daily or longer accumulations.

The Bowen Ratio is a simplified approach to obtaining E by using the energy balance method (Equation 4) and allows for the direct use of measured water temperatures at Dune Lake (rather than indirect use to estimate net radiation as in the Penman-van Bavel method above). All the same uncertainties are inherent in estimating net radiation, but the errors in measuring vapor and temperature gradients are reduced with this method. The weakness in applying this method to Dune Lake was the assumption of "bulk transfer coefficients" between the water surface and the height of the meteorological instruments rather than the classical gradient measurements at 2 different heights above the water surface (Moore, 1983). Limitations of the Bowen ratio include difficulty in determining subsurface fluxes over water, which are potentially important (Ohmura, 1982). Another potential problem is calculating incorrect signs for turbulent flux during early morning and late afternoon, when temperature gradients are changing rapidly, and during precipitation events when turbulent flux is dampened. A third problem occurs when the Bowen ratio approaches 1, when extremely inaccurate magnitudes of evaporation can result.

The surface energy balance can be written as (Thom 1975):

$$Qnet + Qe + Qh + Qw + Qa = 0 \qquad (4)$$

Where:

Qnet = net radiation (W/m^2)

Qe = latent heat flux (W/m^2)

Qh = Sensible heat flux (W/m^2)

Qw = change in water heat storage (W/m^2)

Qa = advection heat flux (W/m^2)

Since the advection term is negligible compared to the rest of the terms, it can be ignored (Kane *et al.*, 1990). If we assume that changes in water heat storage are negligible as well, that is, all of the radiation goes toward evaporation rather than heating the water, the Equation simplifies to:

$$Qnet + Qe + Qh = 0 \qquad (5)$$

The Bowen ratio, β, is defined as the ratio of Qh/Qe (Bowen, 1926) and solving for Qe yields:

$$Qe = (Qnet) / (1 + \beta) \tag{6}$$

$$E = Qe/L = Qnet / ((1+\beta) / L) \tag{7}$$

Where L = latent heat of vaporization

The lake evaporation season used here is May 15- Sept 30, which typically corresponds to the open-water season. Open water at Dune Lake in spring occurs sometime between May 10- May 20 and freeze up occurs early to mid October. Hourly evaporation was calculated using the Penman-Van Bavel method only for 1997, since it had the best data set available. The hourly data was then summed by month and a seasonal value was calculated (May 15-Sept 30). This data set was then compared to the daily calculations for the same time period (May 15-Sept. 30, 1997).

For each open water season for 1995-1998, daily evaporation was calculated using both the Penman-van Bavel and the Bowen ratio methods but using daily summed or averaged data rather than the hourly (Table 3.3).

Table 3.3. Comparison of methods for calculating evaporation.

	Period	**Penman-vB (cm)**	**Bowen ratio (cm)**
1995	July17-Sept 30	15.4	17.9
1996	May15-Sept 30	38.3	49.6
1997	May15-Sept 30	40.4	51.1
1997 (hourly)*	May15-Sept 30	50.7	
1998**	May15-Sept 30	36.5	46.2

*All are daily sums except where indicated.

* *Wind sensor failed mid-June, 1998, so wind speed was estimated by taking an average of previous season's wind speed.

In 1997, the calculated seasonal value of evaporation using hourly data (50.7 cm) was 25% higher than that based on daily data (40.4 cm) calculated using the Penman-van Bavel method and similar to the value (51.1 cm) calculated using the Bowen ratio method (Table 3.4). These values are higher than others for this latitude calculated for Canada (Morton, 1979) but are within the range estimated for other lakes in interior Alaska (Barber and Finney, 2000; Nakao *et al.*, 1981). I also compared the monthly values for both methods (Table 3.4) to see where the greatest discrepancies occurred. The earlier months of May and June had the greatest differences while September was very similar for both methods. For water budget determinations and hydologic model runs, the Penman-van Bavel (PE) method using daily data was used as a minimum and the Bowen Ratio (BE) used as a maximum value to estimate ranges for evaporation.

Table 3.4. Monthly evaporation averages (1995-1998) comparing the Penman-VB daily and Bowen methods.

Date	Penman-VB (cm)	Bowen (cm)
May 15-30	6.7	9.6
May	10.7	15.9
June	12.4	16.4
July	10.7	13.4
August	5.9	7.3
September	3.4	3.1
May-Sept	43.1	56.1
May15-Sep30	39.1	49.7

The well water-level data (Table 3.5) show increasing distance between the top of the water and the top of the well head over time (Figure 3.3) demonstrating that the groundwater level was decreasing over the period studied. Data on groundwater relative to lake-level are shown in Table 3.6. The long-term trend indicates that both lake-level and groundwater levels were declining simultaneously.

Table 3.5. The depth to the level of water (below well head) in each well from 1995-2001.

Date	South East	West	Rick's	Millers	SW Pit
6/8/95	138	184		161.5	
7/2/96	126	180.5	135	165	142
6/29/97	138	185.8	138	166.3	140.5
10/7/97				180.5	140
5/26/98	141	201	147	206	130.5
7/24/98	144	194	146	179	136
7/6/99	165	206	120.5	189	162
7/5/00	168.5	210	172	205.5	163
10/5/01	183		188		
7/2/96-7/5/00 (Total change)	42.5	29.5	37	40.5	21

Table 3.6. Height (m) of groundwater above/below lake surface.

Well	Jul-96	Jun-97	May-98	Jul-00	Oct-01
SE	0.26	0.17	0.26	0.21	0.22
South Pit	0.48	0.52	0.73	0.63	
West	0.02	-0.01	-0.08	0.08	
Miller	-0.14	-0.10	-0.41	-0.17	
Ricks	-0.01	-0.01	0.00	-0.04	-0.05

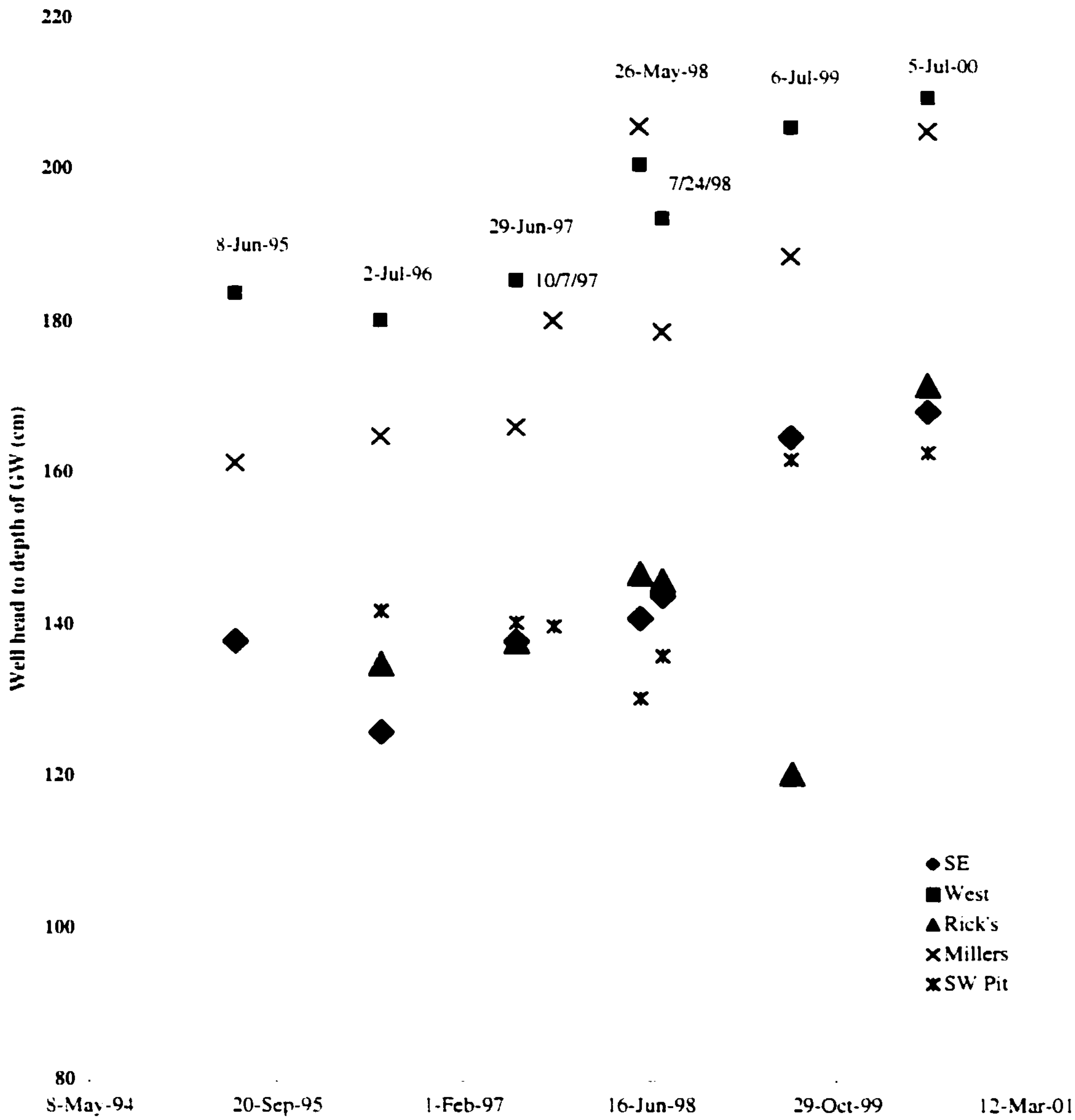

Figure 3.3. Depth of groundwater relative to head at each well.

The groundwater component of the lake water balance was estimated: 1) using the well and lake-survey data and 2) by back calculation using Equation 2. The direction of groundwater movement and the hydraulic gradient can both be calculated from the relative geographic positions of the wells, the distance between the wells and the lake surface and the total head at each well. For this study we used total head at each well as the height of the groundwater above or below the lake-level from survey information. In the first method, hydraulic gradient was calculated from the difference between groundwater and lake surface heights divided by distance between well head and lake at South well for the input and at Miller's well for the output (Table 3.7) (Watson and Burnett, 1993).

The hydraulic gradient (I) is:

I = (Well head (m) – lake surface height (m)) / (Distance between well and lake (m)) (7)

Table 3.7. Hydraulic gradient (m/m) calculated from surveys of groundwater level (GW) relative to lake surface. A positive gradient indicates well-water level above lake-level while a negative gradient indicates well-water level below lake-level.

Date	**SE**	**South Pit**	**West**	**Miller**	**Ricks**
Jul-96	0.020	0.022	0.001	-0.008	-0.001
Jun 30-97	0.013	0.02	-0.001	-0.006	-0.001
May-98	0.020	0.033	-0.004	-0.023	0.000
Jul-00	0.016	0.028	0.004	-0.009	-0.005
Oct-01	0.016				-0.006

The direction of flow was established from the calculation of the magnitude and the sign of I and orientation of the wells. The data clearly indicate the flow of groundwater was from south-southeast to north-northwest. The lake is oriented with its large axis in the approximate north to south direction, so flow comes in from the South Pit well area and moves north toward Miller's well (Figure 3.1.)

These determinations of hydraulic gradient are just snapshots in time and are not necessarily a representative average for the whole open water season. In 1997 we monitored the South Pit well and Miller's well continuously from June 30 to October 6 with well monitoring equipment. Both wells showed decreasing groundwater level over the course of the summer concurrent with lake-level (Figure 3.4). I calculated a daily hydraulic gradient based on the survey data and relative groundwater to lake-level between June 30 and October 6. Miller's well had an average hydraulic gradient of -0.006 ($\pm$0.001) with a minimum of -0.004 and a maximum of -0.007. The South Pit well had an average hydraulic gradient of 0.029 ($\pm$0.002) and a minimum of 0.023 and a maximum of 0.033. Based on this information, the estimate of hydraulic gradient from the initial survey on June 30 was a good average for Miller's well but for South Pit the gradient was underestimated by about 27%.

Once the hydraulic gradient was calculated and direction of groundwater flow established, the ground-water flux was calculated by using the steady state Darcy approximation (Muskat, 1946) which states that flow rate through a porous medium is proportional to the head loss and inversely proportional to the length of the flow path:

$$Q = K * I * A \qquad (8)$$

Where Q = groundwater flux (m^3/d)

K = hydraulic conductivity (est. from grain size) (m/d)

I = hydraulic gradient (m/m)

A = cross sectional area through which groundwater flows

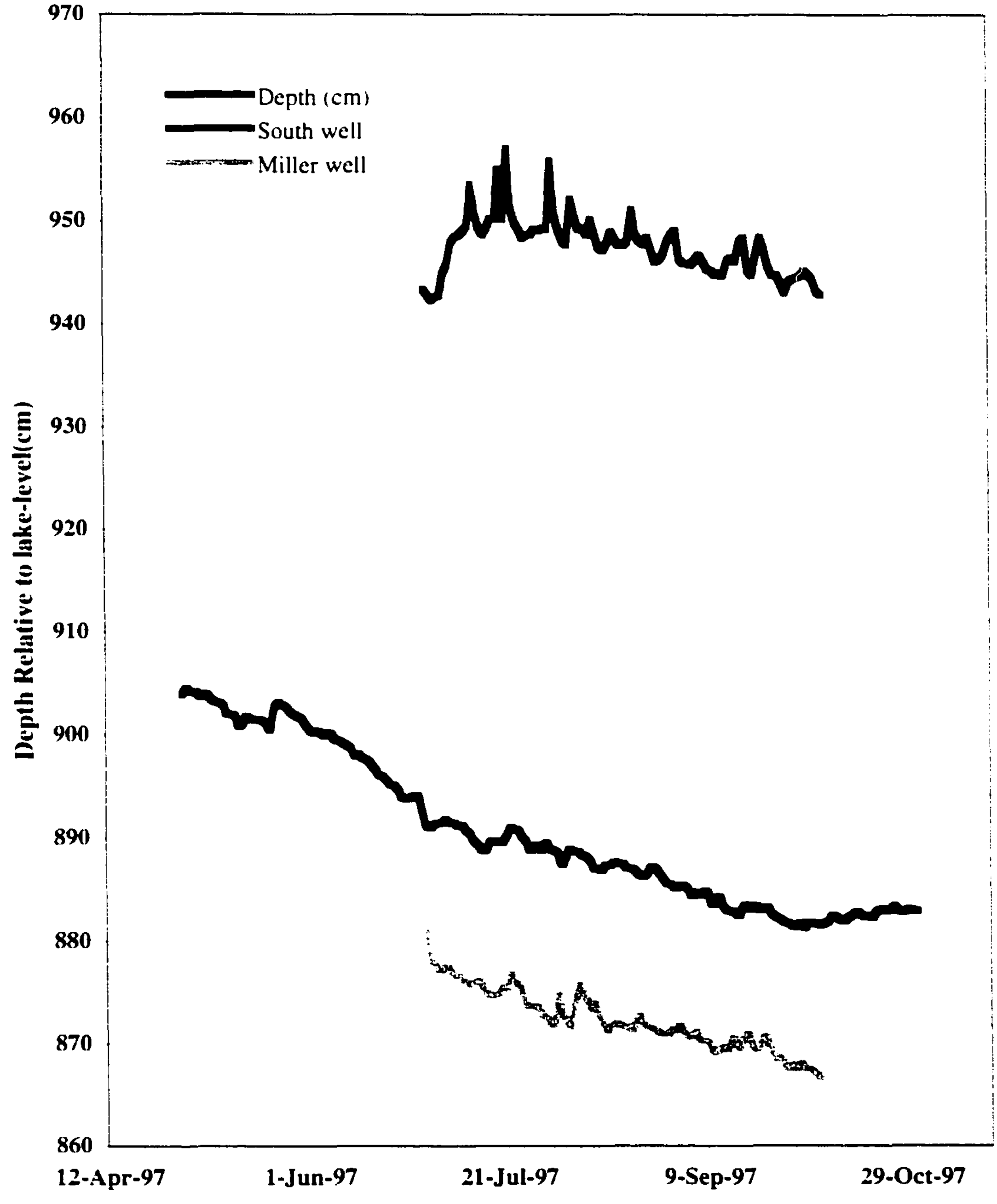

Figure 3.4. Groundwater level at the South Pit and Miller Wells relative to actual lake depth in 1997.

Hydraulic conductivity (1m/day) was estimated from grain size analysis of surface samples, which showed that the surface dune material consisted of 99.3% fine-grained sand. This number might be severely underestimated, by ~10 times, if more of the sand is of medium grain-size. The cross sectional area is an estimate for the size of the aquifer and is equal to the length of the cross-section of shoreline perpendicular to given wells, times the thickness of the aquifer along the cross-section. Since Dune Lake is 9m deep, I used 9 m as the thickness of the aquifer and the width of the lake perpendicular to groundwater flow (650 m) as the width of the aquifer.

The estimates of flux of water into the lake ranged from 126-165 m^3/day and fluxes out of the lake were 35-140 m^3/day (Table 3.8). The flux into the lake was greater than the flux out at all survey times giving an approximate net flux of 53.4 –137.1 m^3/day. This translates to the equivalent of adding about 0.63-1.61 cm of water over the area of the lake between May 15-Sept. 30. These numbers could be up to ten times greater, if the grain-size of the sand is highly variable and is less fine-grained in places other than those measured.

Table 3.8. Groundwater flux into and out of Dune Lake as determined by hydraulic gradient and conductivity. Net seasonal flux is relative to lake surface (1.18 km^2).

Minimum Flux*	**7-96**	**6-97**	**5-98**	**7-00**
Flux in (m^3/d) (South Pit)	126.0	172.0	192.6	165.0
Flux out (m^3/d) (Miller)	49.1	34.6	139.2	55.2
Net flux (m^3/d)	76.9	137.1	53.4	109.7
Net seasonal flux (cm) (May 15-Sept. 30)**	0.91	1.61	0.63	1.29

*Calculations are for hydraulic conductivity of 1 m/d.

**Based on calculation using the hydraulic gradient determined from a single measurement except in 1997 where an average hydraulic gradient was calculated from well data collected between June 30 and September 30.

From the period of about 1995-2000, groundwater declined in the wells an average of about 34 cm which is similar to the lake-level decline. The wells were surveyed at different times of the year making it difficult to compare hydraulic gradients across the years except for 1997 where we had a continuous record between June 30 and September 30.

Another way to estimate groundwater flow is by using a hydrologic model. The model used to determine the water budget of the lake is solved in a spreadsheet (Barber and Finney, 2000). For the open water seasons of 1995 through 1997, lake-level was modeled using calculated evaporation and measured rainfall in Equation 2. The monitoring didn't start until July of 1995 so for this year the entire open-water season is not modeled. The precipitation and wind monitors stopped recording in June of 1998, so that year could not be used for these calculations. Groundwater was specifically left out in the model runs so that difference between modeled lake-level (using evaporation and precipitation as inputs) and recorded lake-level could be used to calculate groundwater.

For each year, the modeled lake-level was lower than actual lake-level (Figures 3.5, 3.6, & 3.7) indicating the importance of groundwater input. The missing component (groundwater) was calculated by difference (Table 3.9). This was done in order to compare this method of estimating groundwater with calculations made from well survey data and hydraulic gradients.

Using Equation 2 and rearranging to solve for groundwater:

$$\Delta G = \Delta V - LA\,(P - E) \qquad (9)$$

Figure 3.5 illustrates actual lake-level recorded by the PT sensor vs lake-level modeled for 1995 using precipitation (P) minus evaporation (E) using Penman-VB (PE) method or using the Bowen ratio (BE). The difference between the actual measured lake-level and the modeled curves is an estimate of the potential net groundwater component.

For the 1995 truncated season, the calculated groundwater is the equivalent of 1.35 cm seasonally (July 17-September 30) or 0.02 cm/day over the whole lake or 236 m^3/day using the Penman-VB method for evaporation. Using the Bowen ratio, the calculated values are 3.64 cm or 0.05 cm/day seasonally or 590 m^3/ day.

The 1996 model run vs lake-level shows that groundwater was a fairly important component (Figure 3.6). Calculating the difference between the curves, the groundwater component is equal to 8.1 cm using PE (0.06 cm/day) and 19.47 cm using BE (0.14 cm/day) seasonally (May 15-September 30) over the lake. This is the equivalent of from 708 m^3/day to 1652 m^3/day.

These calculated values compare less favorably with the calculated groundwater fluxes using the hydraulic gradient method. The calculated range of 708 to 1652 m^3/day is more than 10 times the maximum flux (76.9 m^3/day) estimated from the hydraulic gradient method. This is most likely due to errors in hydraulic gradient and conductivity. In this case, the back calculation of groundwater is probably a better estimate of groundwater.

Estimates of groundwater calculated by difference using the calculated evaporation by the two methods in 1997 (Figure 3.7) ranges from a seasonal average of 1.39 cm (or 0.01 cm/day) for the PE method to 12.47 cm (or 0.09 cm/day) for the BE method. This translates to 118 to 1062 m^3/day or ~ one to nine times the maximum flux calculated from the hydraulic gradient method (137.1 m^3/d). In this case, we used a seasonal average of the hydraulic gradient, (June 30 through Sept 30), so the two methods were comparable. Unfortunately in early June of 1998, wind and rain monitors ceased functioning and not enough data were available for modeling this year.

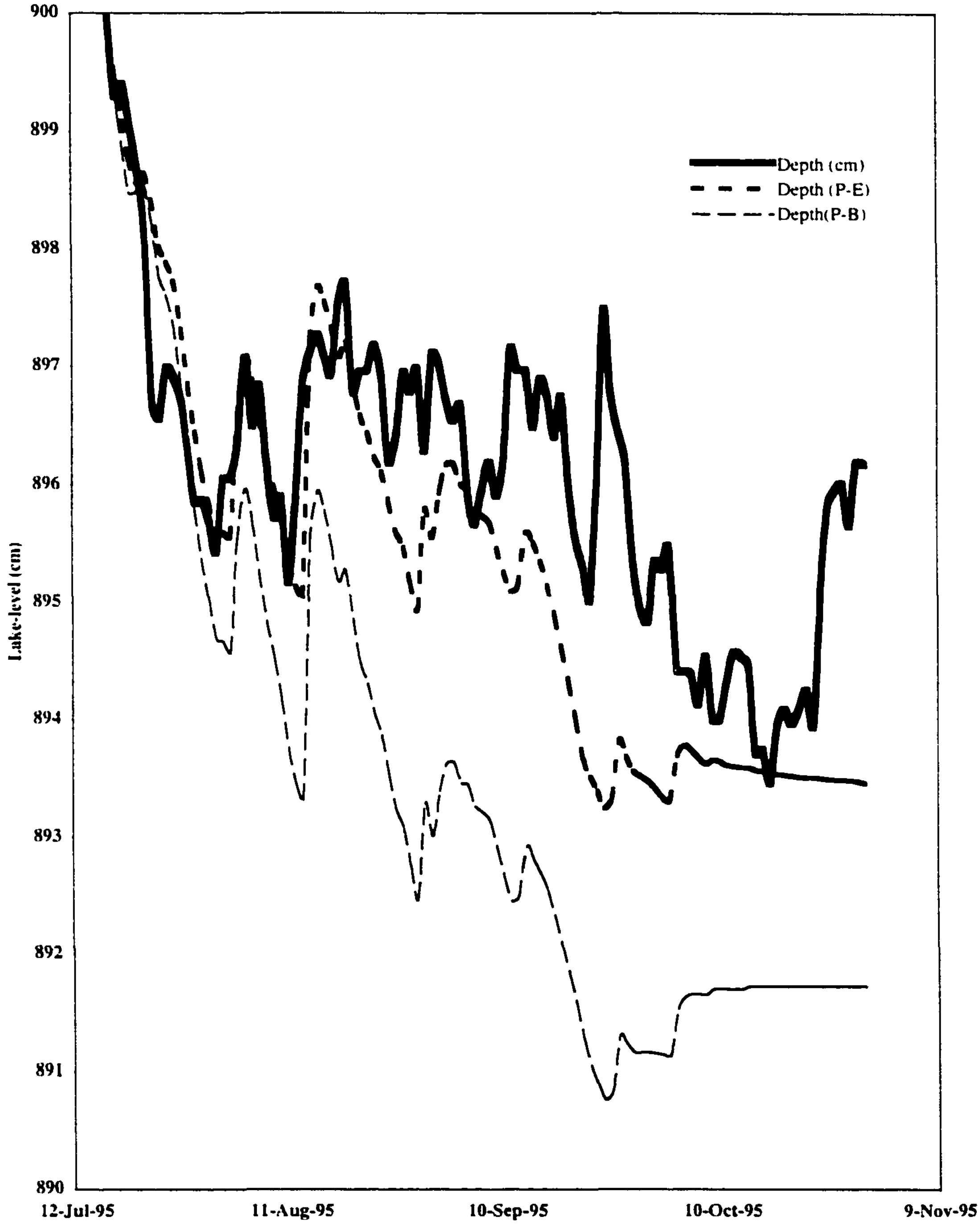

Figure 3.5. Modeled vs actual lake-level for 1995.

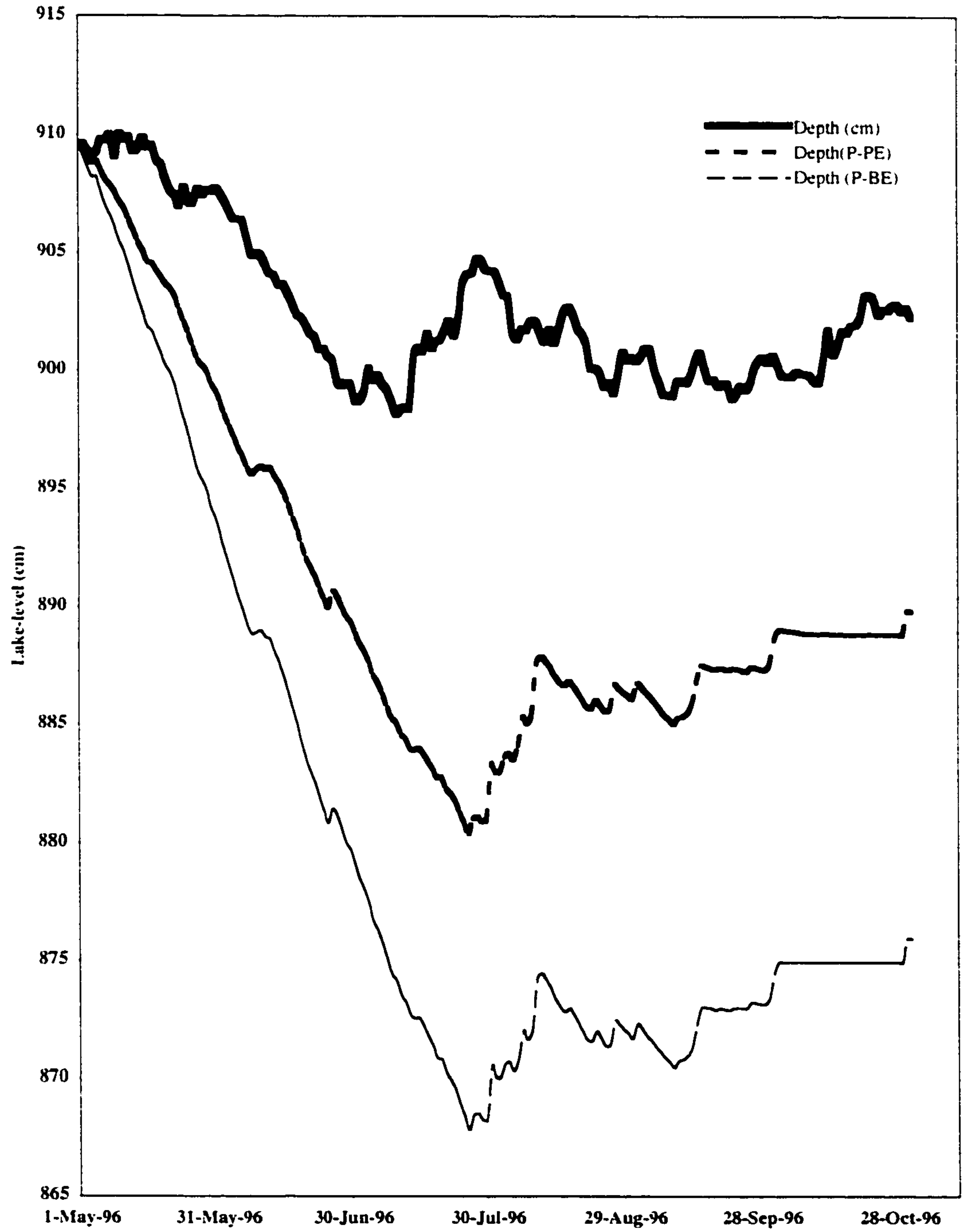

Figure 3.6. Modeled vs actual lake-level for 1996.

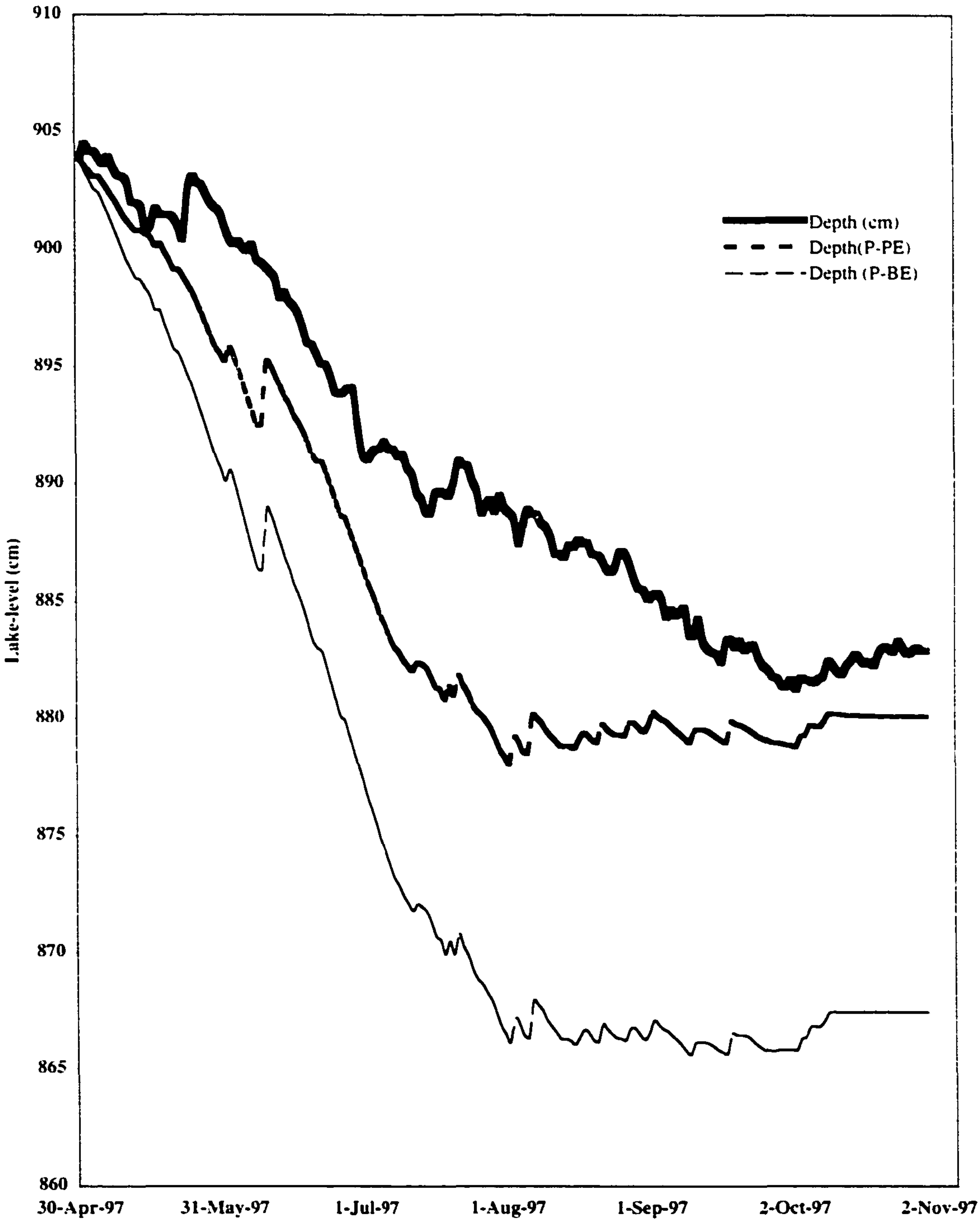

Figure 3.7. Modeled vs actual lake-level for 1997.

Table 3.9. Summary of model parameters for open-water season for 1995-97 and estimates of groundwater flux. All units are in cm and are relative to lake surface area (1.18 km^2).

Year	Lake-level decline	PE	BE	PPT	GW(PE)	GW(BE)
1995*	4.82	15.69	17.93	9.10	1.35	3.64
1996	9.09	38.31	49.57	21.85	8.11	19.47
1997	19.45	40.42	51.07	18.63	1.39	12.47

* 1995 started on July 17

The estimates for groundwater fluxes calculated from well survey data have large errors, not only because of the uncertainties hydraulic gradient and conductivity used, but also because the position of the aquifer depth and width relative to lake-level are unknown. The most reliable estimates for groundwater data are probably those based on the monitored lake-level and calculated evaporation. Based on the 1996 and 1997 data, groundwater input is important to maintaining lake-level, but is smaller than evaporation or precipitation. Groundwater also appears to be much more variable than evaporation.

Discussion

Dune Lake has a dramatic history of recent lake-level changes. Interviews with some local residents indicate a trend of lake-level rising in the early 1970's. Other residents indicate a rapid rise in lake-level starting in the mid-1980's. Lake-level continued to rise until about 1996 after which it started dropping again. We documented peak lake-level mid-May in 1996 with our PT sensor but observations by the cabin owners of one cabin on Dune Lake indicate May 1995 lake-levels were even higher (our PT sensor didn't get deployed until July of 95). Lake-levels continued to decline until the time of our last measurement (October 2001). Several measurement techniques indicate a 60-cm drop in lake-level over the 1995-2001 time period with the biggest change (about 35-40 cm) occurring between July 1999 and October 2001. One of the cabin owners has been documenting lake-level by measuring the distance between the bottom of the cabin and the lake surface since 1995. We calibrated this with our PT sensor data and well survey data and developed a record of lake-level decline for 1995-2001 (Figure 3.8).

Dune Lake is a complex hydrologic system. Lake level is likely driven by groundwater input as groundwater levels have declined concurrently along with lake-level throughout the monitored period (1996-2001). Although the estimates of groundwater are less than 50% lower than evaporation, the input of groundwater is important for maintaining the lake-level. In an effort to determine why groundwater might be declining, I looked at other records collected in Alaska, such as stream gauge and snowpack data. Unfortunately hydrologic and meteorologic records are sparse and of short duration in Alaska.

Precipitation is an important component of the Dune Lake water balance and can be highly variable from year to year. A comparison of the Dune Lake summer precipitation data with data from Fairbanks and Denali Park meteorological station, the only high altitude site in the Alaska Range (Table 3.10), shows similar trends with a small peak of rainfall in June and the major peak in August. Dune Lake had overall higher precipitation than Fairbanks but less than Denali. The average 1996-97 Dune Lake May-August precipitation was 20.7 cm while the average at Fairbanks was 14 cm and at Denali Park was 25.4 cm.

Winter precipitation and snowpack levels could be important to Dune Lake-level, but were not directly measured. By looking at long-term mean of summer precipitation as a percentage of annual for Denali Park and Fairbanks, we can estimate annual and winter precipitation for Dune Lake. Fairbanks winter precipitation is about 37% of the total annual precipitation (long-term mean) or about 10 cm a year. Denali winter precipitation is about 12 cm or 32 % of annual precipitation. Summer precipitation for Dune Lake was 22.05 and 19.35 cm (average 20.7 cm) for 1996 and 1997. Using 65% as an approximation of the percentage of summer to annual yields an annual precipitation of approximately 31.8 cm and winter precipitation of about 11 cm for Dune Lake.

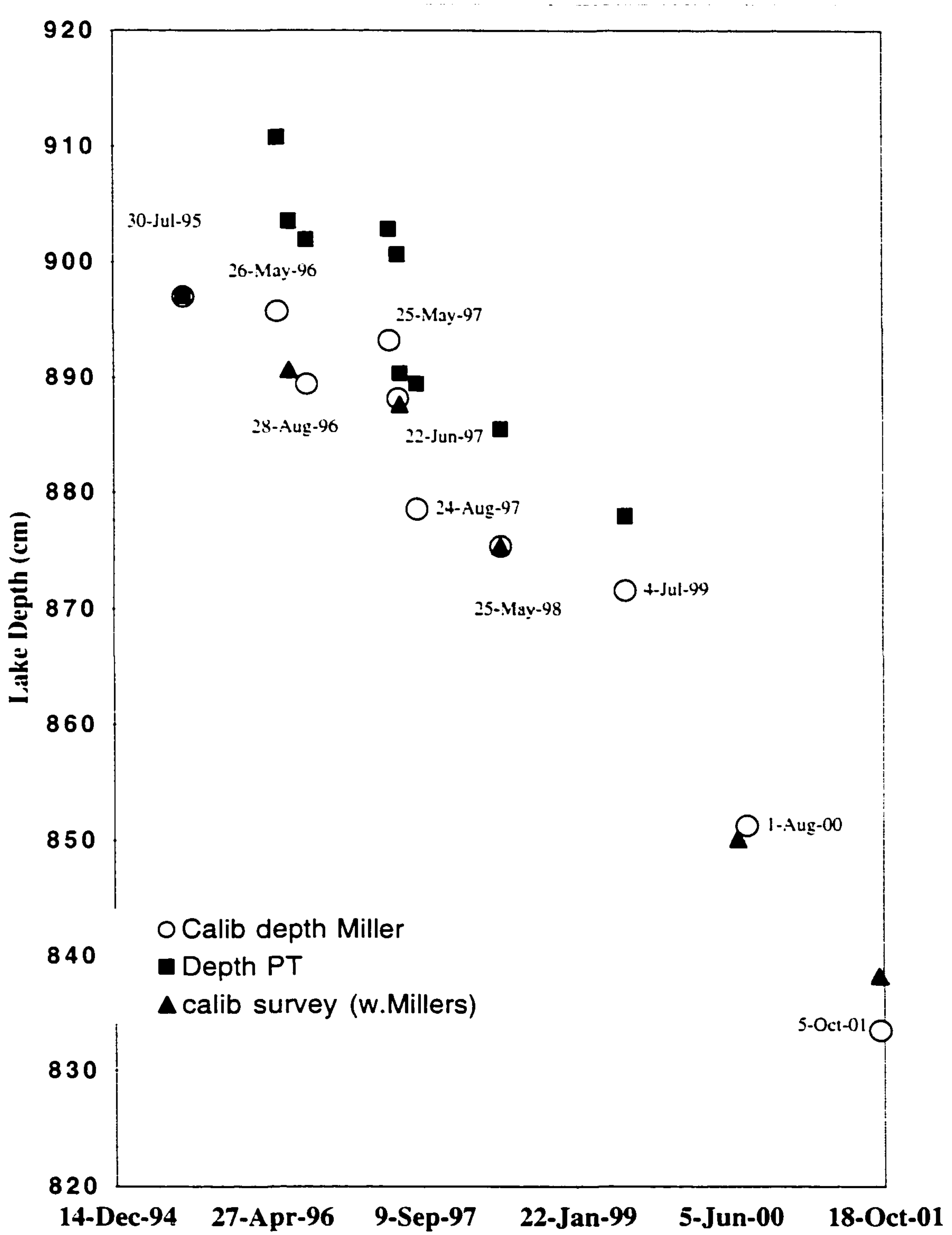

Figure 3.8. Three independent determinations of lake-level calibrated to actual lake-level at beginning of measurement period

Table 3.10. Monthly summer precipitation for Dune, Fairbanks and Denali (1995-1998).

Month	Dune	Fairbanks	Denali
1995			
May		1.85	3.56
Jun		4.85	8.51
Jul		3.35	5.89
Aug	6.40	5.33	5.89
Sep	2.45	3.38	3.38
Total		18.77	27.23
1996			
May	0.50	0.36	1.78
Jun	2.28	3.96	3.33
Jul	5.28	2.72	4.09
Aug	9.55	7.19	8.59
Sep	4.45	2.69	2.67
Total	22.05	16.92	20.45
1997			
May	1.25	0.18	3.86
Jun	5.10	2.62	5.46
Jul	3.08	2.74	7.70
Aug	7.38	4.32	9.35
Sep	2.55	1.22	3.89
Total	19.35	11.07	30.25
1998			
May	1.43	1.04	1.50
Jun	1.30	3.38	4.24
Jul		8.51	11.51
Aug		8.08	8.76
Sep		3.02	3.78
Total		24.03	29.79

There are some patterns in the Denali precipitation record that are probably relevant to the hydrology at Dune Lake. Although the record is missing some data, the trends follow the Fairbanks and McGrath data records (Glenn Juday -personal communication). From the Denali record (Figure 3.9a), we see that growth year precipitation (Sept-Aug) was on the average high between 1962 and 1971. Precipitation dropped from 1971 to its lowest value in 1976. From about 1977 there was a trend of increasing precipitation until 1994, after which precipitation started to decline. This record has similar trends as our Dune Lake level record. We know lake-level was rising from at least the mid 1980's through 1995. Lake-level and precipitation at Denali Park have both declined following this time period.

If we look at seasonal trends in Denali record precipitation (Figure 3.9b and 3.9c), we see other patterns in the record. Summer precipitation (May-August) has been low or about average since about 1991. Winter precipitation (September-April) was low from about 1973 until 1987, after which it increased substantially until 1994. Then it began to decline once again. Between 1994 and 2001, winter precipitation was below the long-term mean 6 out of the 7 years. Looking at the Fairbanks record, summer precipitation has declined over the 96-year record. Except for a couple anomalously high snow years in the early 1990's (1991 and 1993), winter precipitation has been fairly low since the mid-1970's. Winter precipitation declined substantially since about 1994 and through 1999 and some of the lowest snowfall was recorded (since about 1906). The winters of 1998 and1999 were two of the lowest recorded snowfall years in the whole record. Winter precipitation is important to replenishing aquifers in Alaska as snowmelt occurs quite rapidly in spring, delivering a large pulse of water. Since the ground is normally frozen when much of the snow melts, most water is delivered as surface runoff, but this is also a function of the previous fall's soil moisture.

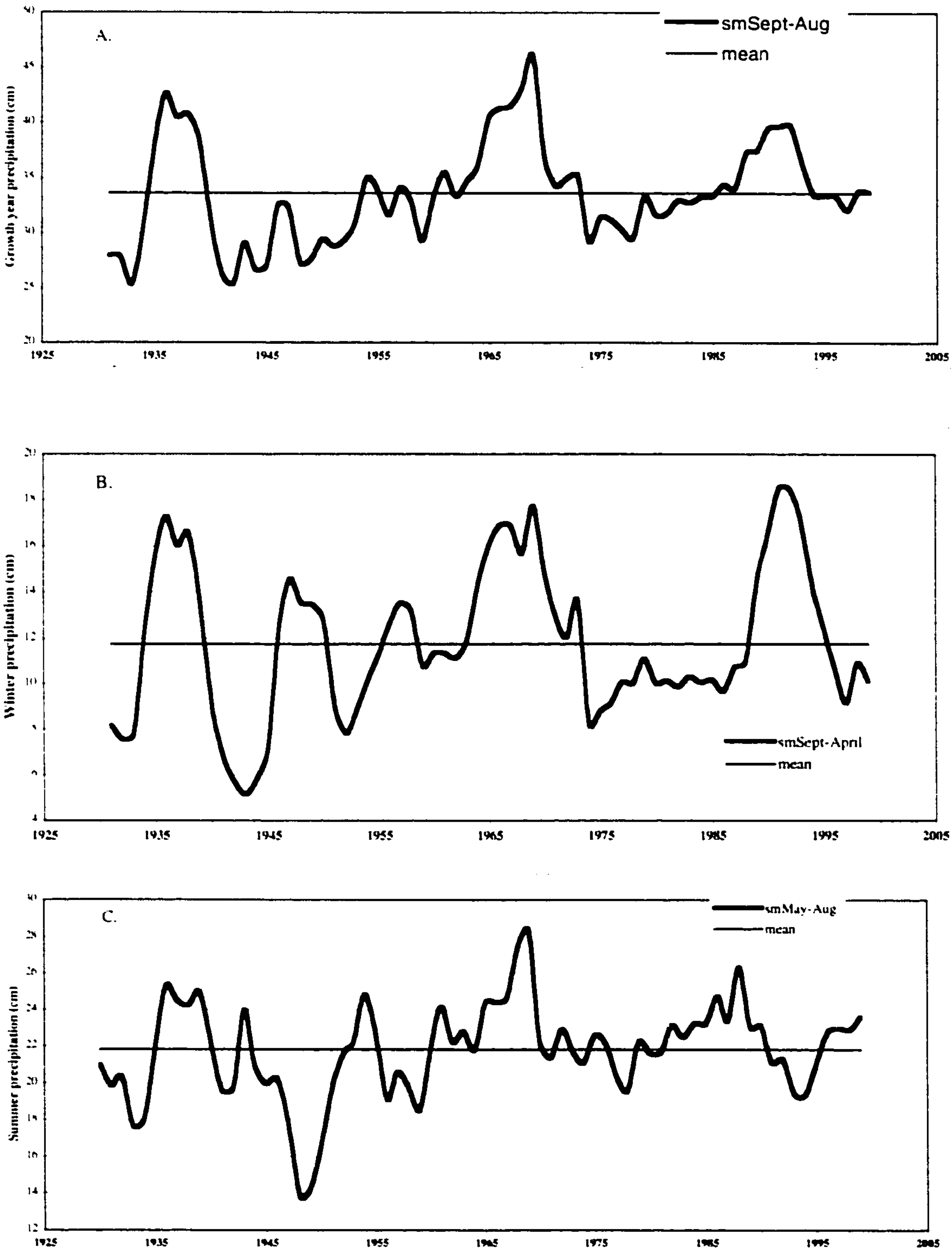

Figure 3.9. Denali National Park precipitation. (A) Growth year (Sept-Aug). (B) Winter (Sept-April). (C) Summer (May-Aug).

Stream flow data from nearby rivers show similar trends. The only long-term record of stream flow in this region (1962-present) comes from the Tanana River at Nenana (Figure 3.10). The Tanana River is 824 km long from its headwaters (confluence of Nebesna and Chisana Rivers) near Northway in eastern Alaska to where it joins the Yukon River. It is 574 km from the river source to Nenana. Approximately 85% of total discharge comes from sources in the Alaska Range, with 50% contributed by 4 glacially fed tributaries (Nebesna, Delta, Nenana and Kantishna Rivers (Anderson 1970)). The peak months of discharge are June-August, when summer melt of snowpack at high altitudes is at its peak and when summer precipitation is at its maximum (July-Aug).

I graphed June-Aug average discharge from the Tanana at Nenana (Figure 3.10), and divided the record into 4 periods; 1962-68, 1969-80, 1981-95 and 1996-2000 (Table 3.11), based on the relationship of average June-Aug discharge (for each period) to mean discharge (1548.6 m^3/s) for this period.

Table 3.11. Mean June through August discharge of the Tanana River by period and compared to long-term average.

	1962-68	**1969-80**	**1981-95**	**1996-2000**
Avg. June – Aug Discharge (m^3/s)	1767.0	1436.5	1593.0	1411.4
Difference from long-term June – August average	+14 %	-7 %	+3 %	-9 %
Minimum difference	-8 %	+6 %	-9 %	5 %
Maximum difference	+42 %	-23 %	+ 17%	-29 %

Summer discharge of the Tanana at Nenana was high during the 1962-1968 period averaging 14 % higher than the long-term mean (1962-2000). Discharge was at its maximum for June-Aug in 1962 and 1967. The flood of 1967 was caused by the highest recorded discharge (2781 m3/sec) in August. Between 1969-1980, discharge was lower than the long-term average by about 7 %. From an Alaska Fish and Game survey in the 1973 and from talking to cabin owners at Dune Lake, we know that Dune Lake was low relative to recent levels in the mid-1970's through the early 1980's. Dune Lake began to rise sometime in the early 1980's (Rick Gold personal communication) and continued to

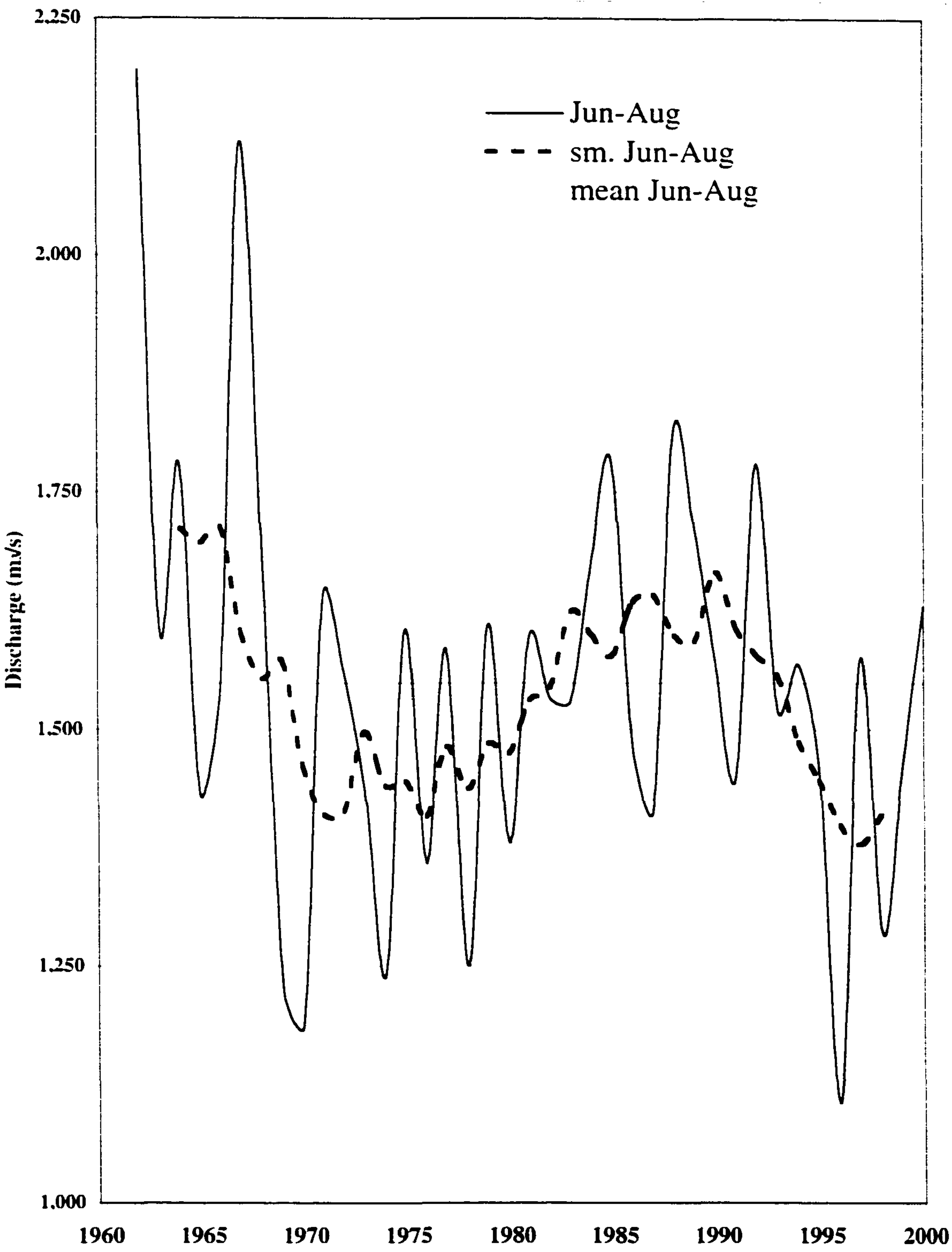

Figure 3.10. Summer (June-Oct) discharge of the Tanana River at Nenana. Annual, 5 yr smoothed and mean values.

rise to a peak in 1995. The summer river discharge during the coincident period (1981-95) was on average about 3 % higher than the long-term summer average and as much as 17 % higher in a single year. Since 1995, lake-level has continued to drop to the present. River discharge decreased after 1995 and from 1996 –2000 was on average 9 % lower than the long-term mean.

Snowpack information was not collected at Dune Lake but we can see some interesting and possibly relevant patterns if we look at snowpack data from nearby areas. Figure 3.11 shows snowpack data collected by USGS from Lake Minichumina to the south of Dune Lake, Farewell, and Bonanza Creek north of Dune Lake. There are many gaps in the data but there are some notable trends. There were a few anomalously high years in the early 1990's, a very low year in 1994 and then low years following 1995 and up to the present. The high snowpack years may have contributed to the rising groundwater and lake-level seen at Dune Lake until about 1995, after which snowpack declined along with groundwater and lake-level.

Thus it appears that Dune Lake-level and groundwater level track river discharge and high altitude annual and/or winter precipitation to a degree. We would expect to see this relationship since aquifers in the region are recharged from summer precipitation and the runoff of high altitude snowpack in spring and glacier melt in summer (personal communication- Bob Burrows, USGS Hydrologist). Summer precipitation consists of 60-70 % of the annual precipitation from the Fairbanks and Denali records and generally falls later in the summer (late July-Aug) when it recharges soil moisture and aquifers depleted from a dry May-June. River discharge also peaks in July-Aug.

The fire in 1981 that burned 720,000 acres occurred mainly to the south of Dune Lake. Loss of vegetation after fire reduces evapotranspiration rates by up to 40% and the burned area takes 15-20 or more years to recover (Calder, 1990). As ET is a major component of water loss in lake water budgets (Barber and Finney, 2000), decreased evapotranspiration rates could cause the lake-level to rise by increasing the proportion of summer precipitation that is incorporated into the ground-water system upstream of Dune Lake. We would expect to see the greatest evidence of this in the several years right after

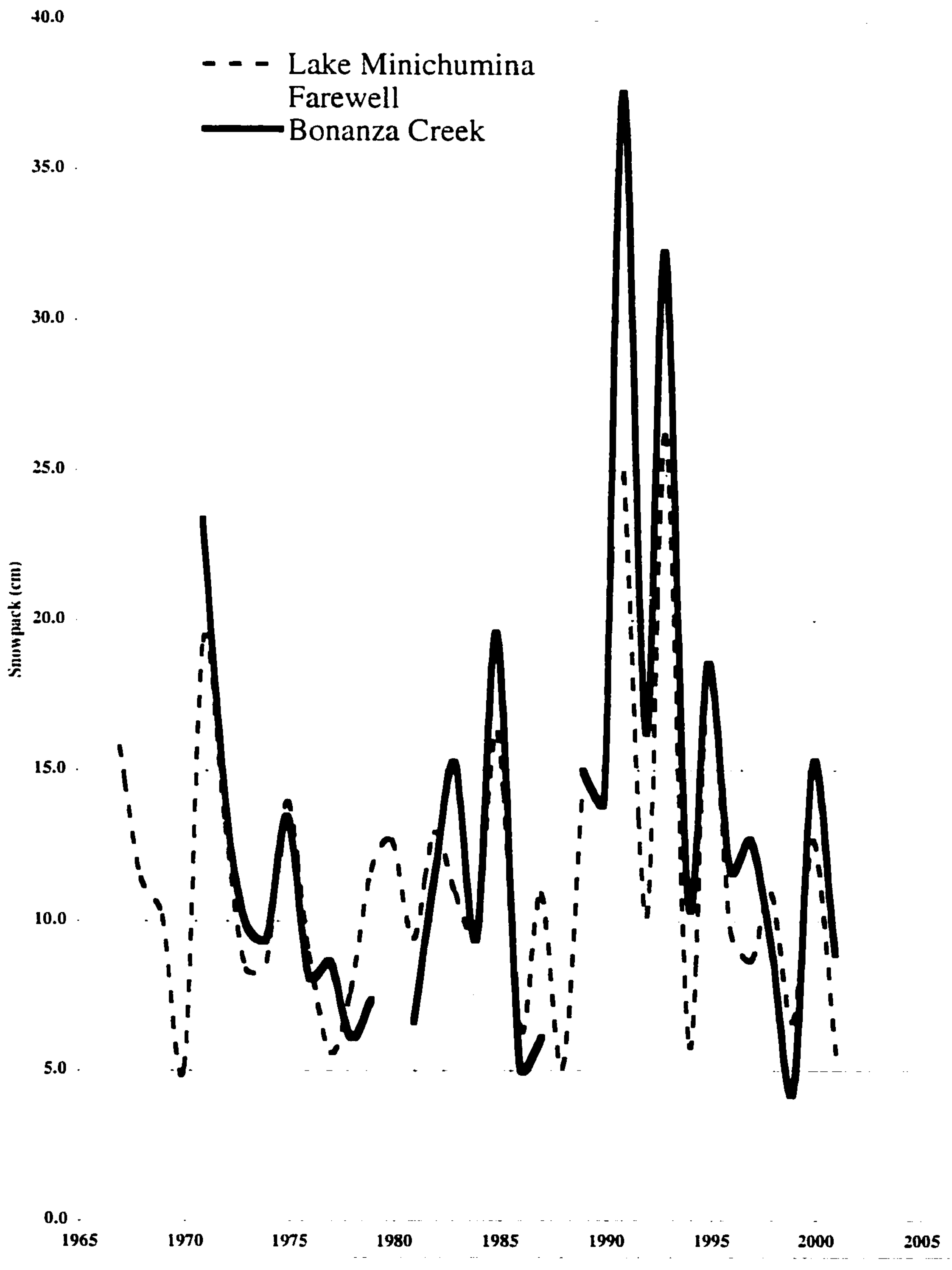

Figure 3.11. Snowpack data from three areas around Dune Lake.

the fire, with a diminished effect over time. In fact, lake-level kept rising until about 1995 when it reached a peak. Thus it appears that the fire did not have a primary effect on the groundwater or lake-level. Disturbance by fire can also affect permafrost, causing the initiation of thermokarst. This could affect groundwater flow in ways that are hard to determine, but a likely effect would be increased drainage from a water table perched on permafrost (Johnson, 1992; Juday *et al.*, 1998; Osterkamp, 1996). If this had occurred, the net effect would have been to reduce groundwater input to Dune Lake. However, as noted, the lake-level rise is inconsistent with this mechanism.

Another unknown factor with regard to groundwater flux history is the relationship to other factors influencing changes in permafrost underlying the area. In other areas of interior Alaska, thermokarsting has been occurring at an accelerated pace over the past 20 years or so (Jorgenson *et al.*, 2001; Osterkamp and Romanovsky, 1999) likely due to increases in annual temperatures. While there is no evidence of ice wedges in the sand sheets (Collins, 1985), surveys have not been conducted. If permafrost is extensive under the sand sheets, thermokarst could cause the water table to drop drastically and the lakes to dry up. There is evidence of dried out ponds throughout the region now but this could also be due to the precipitation and groundwater changes discussed earlier.

Conclusion

While local evaporation and precipitation appear to be important relative to groundwater input at Dune Lake, lake-level reflects changes in groundwater. Changes in groundwater can result from changes in many factors. The fact that 85% of Alaska has permafrost, adds complexity. Dune Lake was at a fairly low level (6 m depth) relative to today in 1975 when it was first surveyed by Alaska Fish and Game as a potential sport fish lake. Stream gauge data of rivers in the interior part of Alaska prior to 1975 show a steadily decreasing trend from a record high, until sometime after the regime change (1977). After 1997 river volumes started increasing. Interior Alaskan temperatures warmed around this time, as a climate regime shift occurred (Ebbesmeyer *et al.*, 1990) and river discharge started increasing with the warmer temperatures, higher snowpack and melting glaciers. The fire of 1981 occurred at a time when lake-level was probably starting to rise from the other factors. The early 1990's also saw some of the highest snowpack levels in about 30 years. Dune Lake continued to rise until it reached a modern peak of about 9 m deep in 1995. Ages of drowned trees indicate that lake-levels were at the highest levels at this time for at least the last 40-50 years. Then lake-level started to decline and has dropped about 35 cm in 5 years. Snowpack levels have declined over the past 5 years, stream flows have declined, the vegetation appears mostly recovered from the fire, and summer temperatures have leveled off and even dropped the past few years. All of these factors together drive groundwater and lake-level at Dune Lake.

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Appendix A:
Lake characteristics, hydrologic parameters.

Lake	Birch Lake	Jan Lake	Dune
AL (km^2)	3.01	0.143	1.18
DA (km^2)	37.00	0.643	6.5
Volume (10^6 m^3)	18.91	0.85	5.09
Mean depth (m)	6.28	5.95	4.76
Median depth (m)	8.34	5.80	4.31
Maximum depth (m)	14.0	12.0	9.0
Mean annual temp. (°C)	-3.44	-5.4	-4.1
Mean ann. Precip. (mm)	328	250	300
Mean ann. E (mm)	600	550	500
Mean ann. ET (mm)	261	183	280
PET (mm)	449	425	440
Discharge (10^6 m^3 yr^{-1})	1.7	none	none
Groundwater(10^6 m^3 yr^{-1}	none	none	0.2
Weather station used	Eielson (1944-1973)	Tok (1954-1984)	

GW flux – 10-350 m3/d

Residence time = lake vol/outflow rate = 5.09x106/ 10-350 =39 –1394 years

[1]Chapter 4

Reduced growth in Alaskan white spruce in the 20th century from temperature-induced drought stress

Abstract

The extension of growing season at high northern latitudes seems increasingly clear from satellite observations of vegetation extent and duration [1, 2]. This extension is also thought to explain the observed increase in amplitude of seasonal variations in atmospheric CO_2 concentration. Increased plant respiration and photosynthesis both correlate well with increases in temperature this century and are therefore the most probable link between vegetation and CO_2 observations [3]. From these observations [1, 2], it has been suggested that increases in temperature have stimulated carbon uptake in high latitudes [1, 2] and for the boreal forest system as a whole [4]. Here we present multi-proxy tree-ring data (ring-width, maximum latewood density, and carbon-isotope composition) from 20 productive stands of white spruce in interior Alaska. The tree-ring records show a strong and consistent relationship over the past 90 years and indicate that, in contrast with earlier predictions, radial growth has decreased with increasing temperature. Our data show that temperature-induced drought stress has disproportionately affected the most rapidly growing white spruce, suggesting that under recent climate warming, drought may have been an important factor limiting carbon uptake in a significant portion of the North American boreal forest. If this limitation in growth due to drought stress is sustained, the future capacity of northern latitudes to sequester carbon may be less than currently expected.

Body

We assembled the available twentieth century summer climate records from stations representative of the boreal forest region of Alaska. (Fig 1a, 1b). Although the greatest magnitude of high latitude warming has been reported in winter and early spring [5], these Alaska stations show a strong warming trend in the growing season over the past 50 years (Fig. 1a, b). Interior Alaska is semi-arid with potential evapotranspiration equal to annual precipitation at many interior locations [6].

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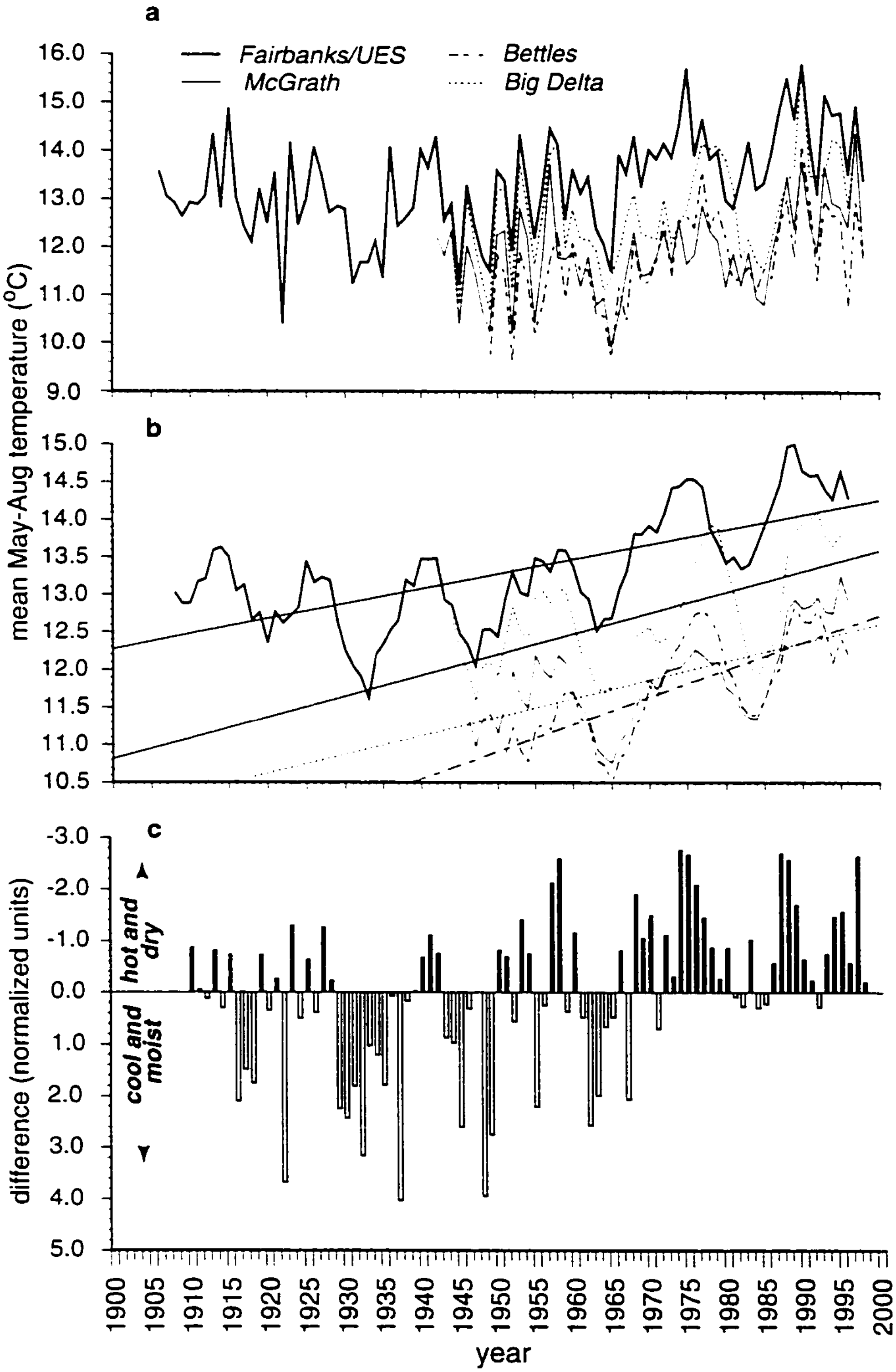

Figure 4.1. Climate trends in interior Alaska in the 20th century.

A combined index of growing season temperature and growth-year precipitation in central Alaska registers severe and prolonged warmth and dryness from the 1970s to present that is unprecedented in the twentieth century (Fig. 1c). Warming without a concurrent increase in precipitation is projected to turn regions of the present-day central Canadian boreal forest into lower productivity Aspen parklands [7].

We sampled trees from a broad range of diameters from closed-canopy white spruce stands in the Bonanza Creek (BNZ) Long-Term Ecological Research (LTER) site and surrounding areas across east-central Alaska (Fig. 2). The sample represents mature and old stands (Fig. 3a) representative of the Alaskan boreal forest, in contrast to the well-studied forest-tundra treeline [8-10].Correlation analyses of all three tree-ring parameters for 36 months (year of tree-ring formation plus 2 years prior) with the Fairbanks/University Experiment Station record show highly significant correlation with summer temperatures (Fig. 4a-c). We did not use other interior Alaska climate stations because they are highly correlated with the Fairbanks record (Figure 1) and are shorter in duration. Ring-width is strongly negatively correlated with summer monthly mean temperature in the year of ring formation plus 2 years prior (Fig. 4a). The negative relationship of ring-width and summer temperature contrasts with the positive relationship widely reported for forest-tundra margin trees [8, 10, 11]. Growth-year precipitation (Sept. through Aug.) is correlated positively ($p>0.01$) with ring-width both for the year of ring formation (0.34 annual values, 0.64 for 5-yr running mean (smoothed)) and the year prior (0.39 annual values, 0.68 smoothed). Most conifers are determinate growers in which photosynthetic gain in the growth season prior to ring formation has the strongest influence on current-year growth [12]. By contrast, the $\delta^{13}C$ and maximum latewood density (MLWD) properties in spruce ring-width are correlated most strongly with mean monthly summer temperatures of the current growth year (Fig. 4b, c).

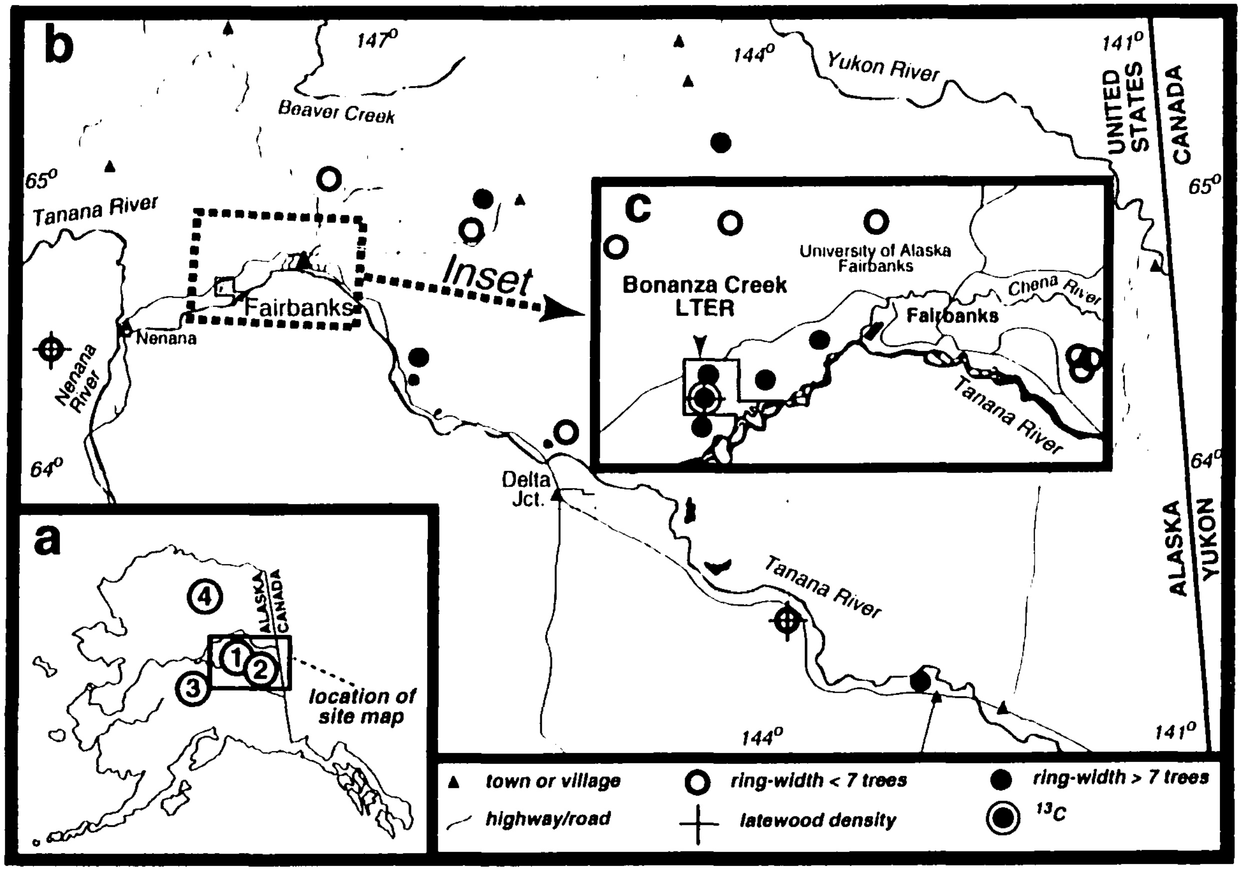

Figure 4.2. Map of field area. (a). Location of meteorological stations. Numbered circles represent meteorological stations: 1 = Fairbanks/University Experiment Station, 2 = Big Delta, 3 = McGrath, 4 = Betties. (b). Extensive tree-ring sampling sites in east central Alaska. (c). Inset of intensive tree-ring sampling locations in and nearBonanza Creek LTER.

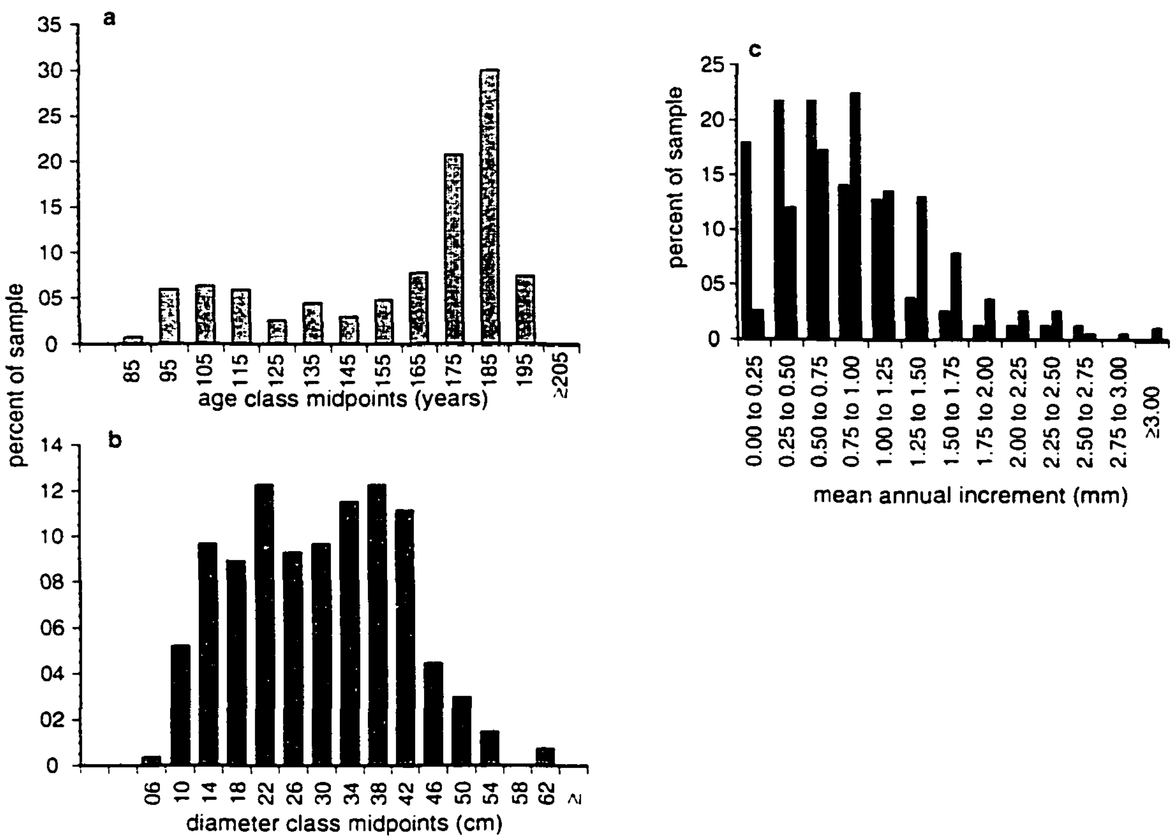

Figure 4.3. Characteristics of white spruce radial growth sample. Minimum tree age (a) and diameter (b), n = 269 trees. c, Frequency distribution of 20th century radial growth for trees correlated ($p<0.05$) with Fairbanks ring-width climate index (n = 191) (black bars) and trees not correlated ($p>0.05$) with ring-width climate index (n = 78) (gray bars).

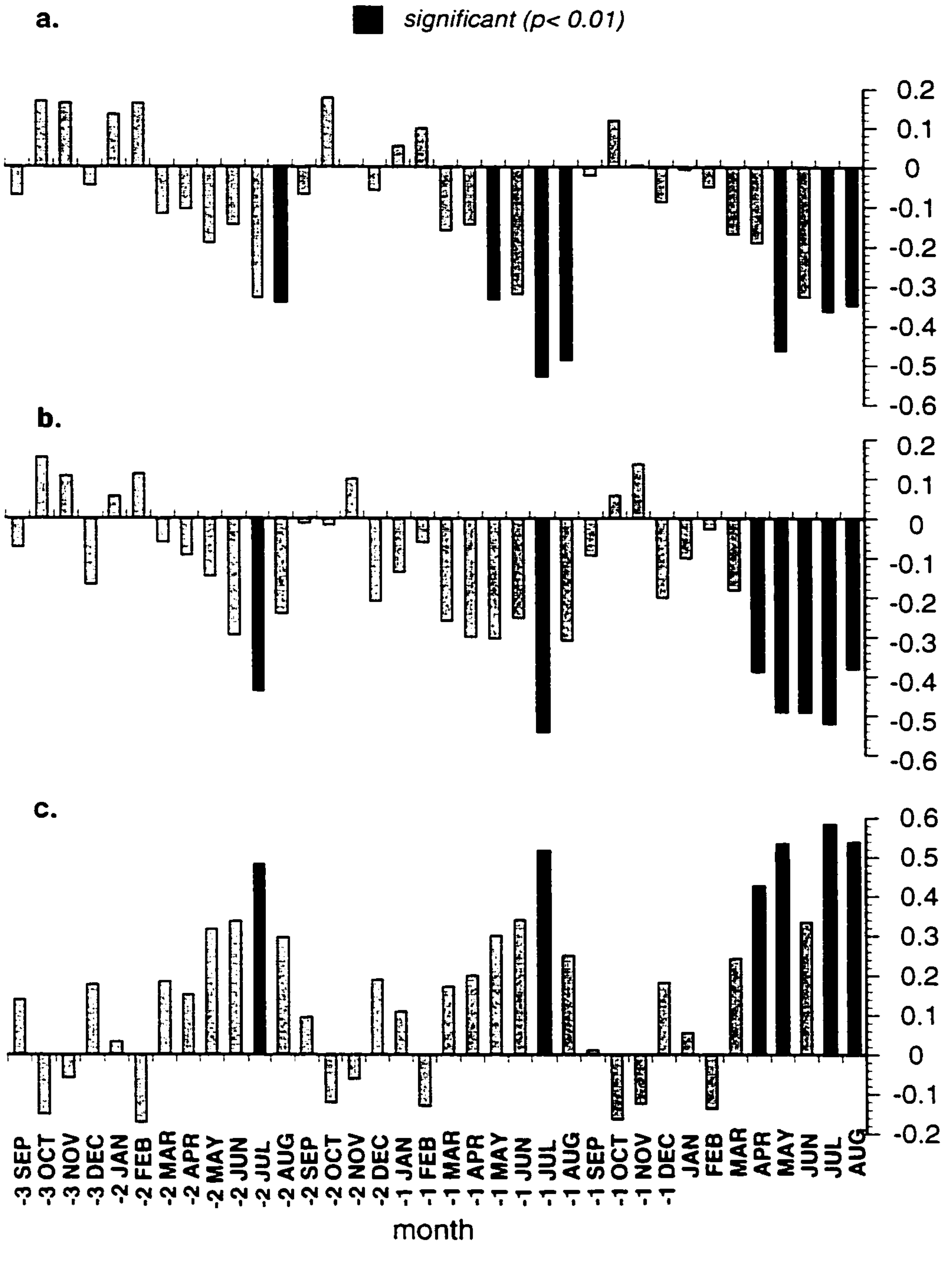

Figure 4.4. Correlation of tree-ring properties to Fairbanks mean monthly temperature for the three years prior to completion of tree-ring growth. The year is indicated by a number from -1 to -3. All values are Pearson correlation coefficients, correlations significant at p<0.01 are shaded black. a. ring-width; b. d13C; c. maximum latewood density. Correlations values are for the period 1909-1996 (a and b), and 1918-1994 (c)

We combined monthly mean summer temperature and growth year precipitation into an index of climatic favorability for white spruce growth (see Methods). High radial growth and favorable climate occur consistently during the 50-yr period 1915-1965 (Fig. 5a). The ring-width climate index (CI_{rw}) closely mirrors one- and two-year positive and negative growth anomalies in the sample (Fig. 5a), such as the spike of growth in response to a cool and moist 1937 and the short, sharp reduction of growth in response to the warm and dry years of 1957-58 (Fig. 1c). A quasi-decadal cyclicity is apparent in which opposite trends in mean radial growth occur regularly over periods of 5-6 years (Fig. 5a, b). The most striking feature of the climate and ring-width record is the sustained unfavorable climate signal and growth response from the 1970s to the present.

Several features of the sample indicate that this climatically related growth reduction is significant at the landscape level of carbon uptake. Mean ring-width of each of the 20 stands and radial growth of a majority (71%) of the individual trees in the sample are significantly correlated ($p>0.05$) with CI_{rw}. The population of trees that is significantly correlated with climate displays a distribution skewed towards faster rates of growth in the twentieth century than the portion of the population that is not significantly correlated (Fig. 3c). Thus the fastest growing trees are most limited by summer warmth and low precipitation.

Correlation analyses of the $\delta^{13}C$ and wood density properties provide independent confirmation of the ring-width results. Carbon isotope ($\partial^{13}C$) ratio provides information on CO_2 uptake and water vapor loss during photosynthesis [13], and thus under limiting conditions can register drought stress. We calculated discrimination of $\delta^{13}C$, the difference between $\delta^{13}C$ of the tree-ring (hollocellulose) and atmospheric $\delta^{13}C$ annually. The $\delta^{13}C$ discrimination of our sampled white spruce is significantly negatively correlated with summer monthly mean temperatures in the year of the ring formation (Fig. 4b).

Fairbanks precipitation is significantly correlated with $\delta^{13}C$ discrimination (0.335, $p>0.02$), but combining summer temperature with growth year precipitation did not result in a significantly higher overall correlation than the correlation with summer temperature alone. The $\delta^{13}C$ discrimination of white spruce displays a strong relationship to the short-

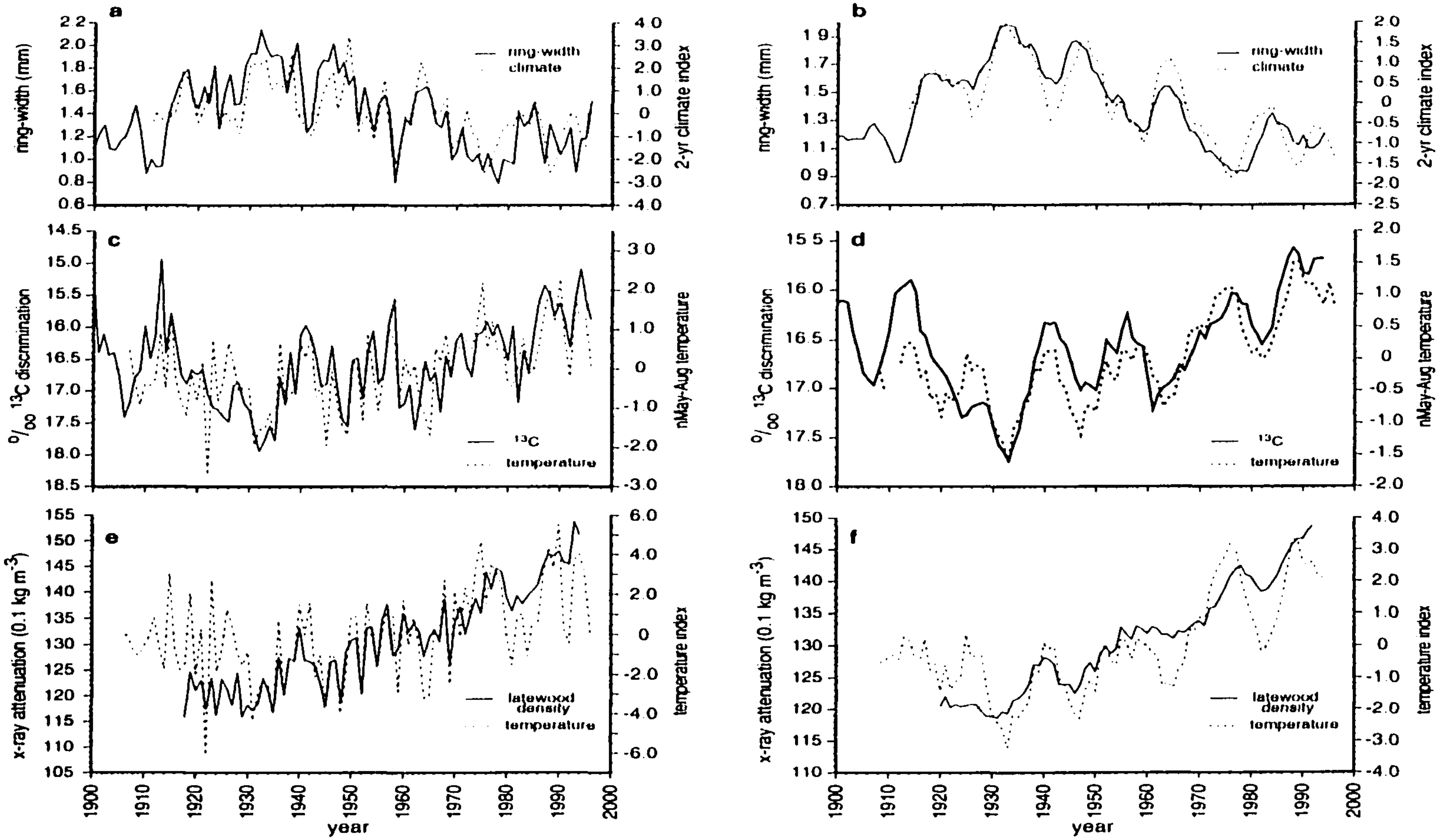

Figure 4.5. Tree-ring properties in relation to Fairbanks climate during the twentieth century. Annual values on left, smoothed (5-year running mean) on right. All climate values are normalized with zero mean and scaled as standard deviation units. All correlations are significant at p<0.001. (a) ring-width versus 2-year average Fairbanks climate index (see text for definition of ring-width climate index, see Fig. 1c for 1-year values of ring-width climate index), correlation = 0.75, (b) smoothed ring-width and climate index, correlation = 0.91, (c) d13C versus May through Aug. temperature, correlation = 0.69, (d) smoothed d13C and May through Aug. temperature, correlation = 0.85, (e) maximum latewood density versus density climate index (mean of normalized May, July, and Aug. temperature), correlation = 0.75, and (f) smoothed maximum latewood density and density climate index, correlation = 0.92.

term variability of Fairbanks summer temperature throughout the twentieth century (Fig. 5c) and declines in the 1980s and 90s to an unprecedented and sustained low level for the twentieth century (Fig. 5d).

MLWD is the maximum density of the wood formed at end of the growing season when conifer tracheids switch from large diameter, thin walled cells (formed in spring and early summer) to smaller diameter, thick walled cells. Correlation analysis has shown that this number integrates growing conditions over the length of the photosynthetically active season [14] and is also more consistent than ring-width [15]. We hypothesize that an increase in the length of the growing season along with warmer summer temperatures depletes already limiting soil moisture. In an environment of high soil moisture tension, wood cell production shifts earlier in the summer from relatively low-density earlywood to higher-density latewood, prolonging the wall thickness phase of growth which is associated with increased latewood density [16, 17]. The MLWD of our white spruce sample is significantly correlated with early and late summer temperatures in the year of ring formation (Fig. 4c), as noted for North American treeline white spruce [15], and other months add little improvement. An unprecedented late twentieth century rise in MLWD and associated temperature index is consistent with the other tree-ring properties (Fig. 5e, f).

Our carbon isotope and MLWD results confirm that drought stress accounts for the climate sensitivity of ring-width for productive upland white spruce sites in interior Alaska. Climate sensitivity in our sample has been sustained (neither degraded nor improved) over the last 9 decades, and drought stress has reached levels unprecedented in the twentieth century because of climate warming experienced in the 1980s and 90s. The wide distribution (Fig. 2), representative diameter distribution of the sampled tree population (Fig. 3a, b), and consistent response of the stands suggest that this response is typical of upland white spruce forests in interior Alaska. In the western Canadian boreal forest, white spruce radial growth displays a negative relationship with the duration of fire weather (hot and dry) [18], suggesting that our findings have broader applicability. Our results also are consistent with the suggestion that drought stress is one possible explanation for a late twentieth century decrease in the positive relationship of temperature and radial-growth of trees at the forest-tundra margin across the Northern Hemisphere [19, 20].

White spruce is one of the most productive and widespread forest types in the boreal forest of western North America [21-23]. Thus any coherent, climate-related change in white spruce growth is likely to be an important factor in CO_2 uptake in the boreal forest, a region that is one of the planet's major potential carbon sinks. Our results suggest that the current assumption of carbon-cycle models [24] of a uniform positive relationship of atmospheric carbon uptake to high latitude warming, will lead to an over-estimate of high-latitude carbon storage and an underestimate of future atmospheric CO_2.

Methods:

The radial growth sample is representative of trees in mature and old stands that are dominant on the contemporary landscape, including trees across a broad range of diameters. The radial growth sample includes 269 trees in 20 separate stands; radial growth was calculated as a stand-weighted mean. All trees contributed from 1908 until the date of sampling which took place between 1994-1996 or in 2 cases the death of the stand. One stand (Reserve West in BNZ) was killed by wildfire in 1983 and the other was logged in 1987. In these 2 stands wood disks were harvested from the stumps. Density was measured in cores from 23 trees collected in 3 stands including Reserve West and outlying sites to the east and west (Fig. 2). The common period of analysis for density was 1918-1994; all trees contributed throughout, except sample depth in one stand dropped from 14 to 6 trees between 1982-1994. Stable carbon isotope ($\delta^{13}C$) was measured only at Reserve West. Four orthogonal cores from each of 4 trees pooled together give an isotope value representative of a site [25]. We obtained isotope measurements for the period 1900-1981 from 4 orthogonal wedges from each of 4 harvested stump disks. For the period 1982-1996 isotopes were obtained from 4 cores collected at each of 4 surviving trees near the fire perimeter. We believe the isotope sample is representative of white spruce of the interior Alaskan boreal forest because Reserve West is centrally located, is within the well-studied LTER, and the tree-ring parameters are well correlated with each other. We confirmed the trends in $\delta^{13}C$ seen at Reserve West in limited samples covering 3 and 4 decades collected at 2 other sites located about 30 km from BNZ.

We did not transform ring-widths into de-trended normalized ring-width index values for two reasons. First, we wanted to preserve the weighting effect of tree increments of different sizes, which reflects annual production more closely than normalized values, which remove this effect. Second, most white spruce of the age range in this study exhibit little age-related growth trend, and the removal of trend is user-specified and risks the loss of information on long-term climate variability that may actually influence tree growth [26]. However, we also correlated climate with detrended ring width indices [27] for several stands and found little difference, demonstrating that our results are not an artifact of age-related changes in ring width.

We examined various combinations of climate parameters in order to produce annual climate indices that maximized correlation for each of the 3 tree-ring properties (Fig. 5). The climate index (CI_{rw}) that correlates best with ring-width (Fig. 5a,b) is composed of the following parameters:

$$CI_{rw} = ((Yr_0\,(PPT)_n - (T_{M\text{-}A})_n) + (Yr_{-1}\,(PPT)_n - (T_{M\text{-}A})_n))/2$$

where: Yr_0 = year of ring formation, Yr_{-1} = year prior to ring formation, $(PPT)_n$ = normalized growth year precipitation, $(T_{M\text{-}A})_n$ = normalized May-August temperature).

For stable carbon isotope analysis, 4 orthogonal samples of annual rings were excised, ground, and blended with the same-year wood from 4 tree disks. Hollocellulose was extracted from the annual wood [28] and carbon isotope ratios were analyzed for the years 1900-1981. The same procedure was used on cores from the surviving trees for the period 1967-1996, to provide an overlap period for correction. Isotope trends for the period of 1967-1981 follow similar curves from both sets of wood samples, but there is a slight offset (0.7‰) in the $\delta^{13}C$ of the fire-killed vs. the live trees likely due to site-specific effects. A correction was made for this offset. A correction was also made for $\delta^{13}C$ changes in atmospheric CO_2 over the last 150 years due to fossil fuel and biomass combustion using the Law Dome Antarctic ice core data (1.53‰ over 150 years) and discrimination was calculated [29]. Thus a continuous record of $\delta^{13}C$ discrimination was determined for 1900-1996 and the best correlation was found with May-Aug temperatures (Fig. 4b, 5c,d).

Maximum latewood density was measured by x-ray attenuation [15, 30] at Lamont-Doherty Earth Observatory. From 1 to 4 radial samples were collected from each tree. Annual density values were first combined to produce annual tree averages that were then used to calculate overall yearly stand means for each of 3 stands. The 3 stand means were then averaged to produce the sample mean used in correlation analysis. . We determined that MLWD correlated best with May, Jul and Aug temperatures (Fig. 4c, 5e,f).

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[1]Chapter 5

Reconstruction of Summer Temperatures in Interior Alaska from Tree-Ring Proxies: Evidence for Changing Synoptic Climate Regimes

Abstract

Maximum latewood density and $\delta^{13}C$ discrimination of Interior Alaska white spruce were used to reconstruct summer (May through August) temperature at Fairbanks for the period 1800-1996, one of the first high-resolution reconstructions for this region. This combination of latewood density and $\delta^{13}C$ discrimination explains 59.9% of the variance in summer temperature during the period of record 1906-96. The 200-yr. reconstruction is characterized by 7 decadal-scale regimes. Abrupt regime changes are suggested at 1816, 1834, 1879, 1916, 1937, and 1974, and appear to be the result of synoptic scale climate changes. The overall mean summer temperature for the period of reconstruction was 13.49 while the recorded temperature averaged 13.31 °C; the coldest interval was 1916-37 (12.62 °C) and the warmest interval was 1974-96 (14.23°C) for the recorded data. The reconstruction is anomalous compared to other Northern Hemisphere records, especially because of interior Alaska warm periods reconstructed from 1834 to 1851 (14.24) and from 1862 to 1879 (14.19). Autogenic effect of tree growth on the site, altered tree sensitivity, or novel combinations of temperature and precipitation cannot be ruled out as contributors to the anomalously warm 19th century reconstruction, but do not appear to be likely. The early 20th century cool/wet period was conducive to the highest radial growth and carbon uptake by productive white spruce stands in interior Alaska over a 200-year period.

[1] Submitted: Barber, V.A., Juday, G.P. and Finney, B.P. Reconstruction of summer temperatures in interior Alaska from tree-ring proxies: Evidence for changing synoptic climate regimes. Climatic Change.

Introduction

Understanding how climate variability affects the functioning of ecosystems is of fundamental importance for natural resource management and ecological science. Recent studies of ice cores, lake-sediment, and tree-rings have documented many climate fluctuations during recent millennia. During this time decadal-to-century scale climatic changes were common, especially in the high latitudes, and included notable intervals such as the Medieval Warm Period (Hughes and Diaz, 1994) and the Little Ice Age (Bradley and Jones, 1993; Overpeck *et al.*, 1997; Wiles and Calkin, 1990). Often these climate changes produce a long-lasting ecological imprint in basic ecological functions such as primary production or reproduction of species. Occasionally a suite of proxy data from this climate-mediated pattern of ecosystem change can be assembled to reveal the fundamental pattern of climate variability itself.

A robust result of General Circulation Model (GCM) simulations is that high-latitude land masses in the northern hemisphere will experience the greatest magnitude of warming under scenarios of increased anthropogenically produced greenhouse gases (Houghton *et al.*, 1996). There is growing paleoclimatic evidence that the 20^{th} century was warmer than previous centuries (D'Arrigo and Jacoby, 1993; Jacoby *et al.*, 2000; Jacoby *et al.*, 1996; Mann *et al.*, 1998; Overpeck *et al.*, 1997). Equally impressive is the strength of evidence that the climate in certain high latitude regions has warmed markedly and abruptly in the last part of the 20^{th} century (Chapman and Walsh, 1993; Houghton *et al.*, 1996) and that large-scale ecological changes are already underway (Serreze *et al.*, 2000). For example, strong warming since the 1970s in Alaska is associated with thawing permafrost (Osterkamp, 1996; Osterkamp and Romanovsky, 1999), very large scale insect outbreaks (Werner, 1996), receding glaciers (Wiles *et al.*, 1996) and a decline in the areal extent of Arctic sea ice (Chapman and Walsh, 1993; Serreze *et al.*, 2000; Wadhams, 1995). The area burned annually by wildfire in Canada has increased dramatically since the late 1970s at the same time annual temperatures have warmed (Kasischke and Stocks, 2000; Kurz *et al.*, 1995). It is tempting to interpret recent warming and ecological changes in Alaska as evidence for the greenhouse gas – climate-warming theory, but before doing so it is crucial to know whether other similar rapid climate changes and conditions occurred in the past.

Controls on 20th Century Climate

Interior Alaska is a well-defined, physiographically complex region bounded on the north by the Brooks Range and on the south by the Alaskan Range (63-67°N). Interior Alaska extends eastward to the Yukon Territory and westward to a climatic boundary (140-155°W) where precipitation exceeds 400m (Edwards *et al.*, 2001). The region is made up of large, low-lying tectonic basins (Tanana Valley and Yukon Flats) separated by uplands (500-1000m in elevation). The Brooks and Alaska Ranges act as topographic barriers to moisture-laden air from surrounding oceans. Consequently interior Alaska is semi-arid, has a precipitation range of 400 to <200 mm annually (Patric and Black, 1968). Precipitation generally declines to the east, and is strongly influenced by topography (Edwards *et al.*, 2001). About 60% of the annual precipitation falls as summer rain. The climate is cold continental with January means -20°C or colder and July means 15-20°C (depending on elevation and location within the region).

Unfortunately, climate records in Alaska are relatively sparse. Most date from no earlier than mid-20th century. The oldest continuous record from interior Alaska (Figure 5.1) is a combination of University Experiment Station (UES) and Fairbanks Airport data (Juday, 1984). The UES/Fairbanks record begins in the first decade of the 20th century. Fairbanks data are representative of the climate found in interior Alaska. Average July temperature for Fairbanks is about 16°C over the recorded period (1906 - present), but has risen to about the 17°C when averaged over the last 20 years. Fairbanks annual precipitation is around 28.2 cm and has not increased over the past 20 years.

Interior Alaska has two distinct summer circulation patterns apparent in records of inter-annual climate. Summer climate is affected by mid-tropospheric variations of ridges and troughs with July and August normally the wettest months (Barry and Hare, 1974). Twentieth century summer climate in interior Alaska has alternated between periods of colder and wetter or warmer and drier (Figure 5.1) than the long-term mean (Edwards *et al.*, 2001; Mock *et al.*, 1998).

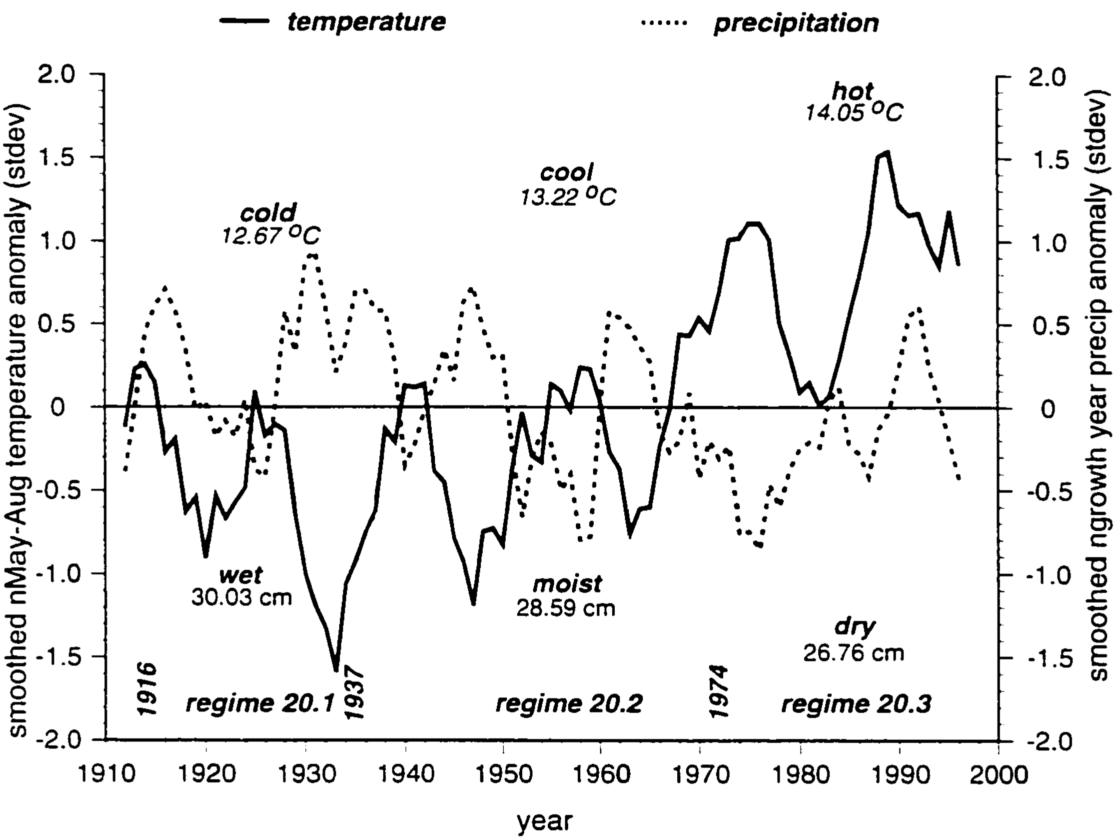

Figure 5.1. Smoothed growth year precipitation versus summer temperature anomalies for Fairbanks recorded data. Data are smoothed to a 5-yr. running mean. Boundaries of summer temperature regimes are indicated, and within-regime means displayed for temperature (above long-term mean or "0" line) and precipitation (below).

The synoptic pattern for colder and wetter conditions is produced by an eastward shift of the East Asian trough and a stronger than normal Pacific subtropical high. Both the eastward shift and the intensification of the subtropical high cause increased frequency of storms containing moisture-laden air to enter the interior basin from the west.

A high-pressure ridge located north to northeast of Alaska acts as a major circulation center and leads to warm and dry summers in the interior. The high center brings clear skies and warm dry continental air from the east at the season of maximum surface heating from the long daylight hours near the Arctic Circle. Negative surface-pressure height anomalies over the Yukon Territory of Canada and over northern Siberia represent a northwest shift of the average pressure system. The Pacific subtropical high located south of Alaska is weak under this regime, resulting in reduced flow of moist air from the west. Persistent blocking ridge conditions are directly correlated with periods of extensive wildfires across the western North American boreal forest (Johnson, 1992).

Circulation patterns associated with: 1) cold and dry or 2) warm and wet summer conditions in interior Alaska are rare. For example, recent studies (Edwards *et al.*, 2001; Mock *et al.*, 1998) were able to find only one example each of warm/wet and cold/dry July climate between 1946-1989. Cold and dry conditions appear to have resulted from a southward displacement of the jetstream as compared to normal with troughing and a westerly wind maximum far south of Alaska into British Columbia (Mock *et al.*, 1998). North of this troughing, colder and drier air masses predominate, a condition that resembles spring circulation. Warm and wet conditions are set up when troughing is centered more westward than normal and ridging is prevalent over Alaska. Convective activity might explain the positive precipitation anomalies during the time that temperature anomalies are high (Mock *et al.*, 1998).

Ecology and climate sensitivity of white spruce

White spruce has a well-defined climate optimum centered on mean July temperatures of 12-15°C and annual precipitation values of 750-1050 mm as indicated by: a) response functions generated by modern pollen studies across the distributional limits of the genus *Picea* (Anderson *et al.*, 1991) and b) the relationship of climate versus modern plant distribution of *Picea glauca* (Thompson *et al.*, 2000). According to these parameters, upland white spruce on low-elevation sites in interior Alaska occupies both the dry and warm extremes of the climatic range for the species.

Most dendrochronological literature on white spruce (*Picea glauca* (Moench) Voss) in western North America references treeline collections (Garfinkle and Brubaker, 1980; Jacoby and D'Arrigo, 1989). In such studies sampled trees were thought to contain a singular climatic signal (usually temperature) and were free from canopy competition. The literature is dominated by studies of trees displaying a positive ring-width response to summer temperature (e.g. (Jacoby and D'Arrigo, 1989). However, in the second half of the 20^{th} century, trees at treeline across high northern latitudes have become less sensitive to temperature at many locations. In such treeline trees, additional summer warmth has produced either limited or no additional growth (Briffa *et al.*, 1998), and lack of moisture may now be limiting growth in contrast to low temperatures (Jacoby and D'Arrigo, 1995). In western Canada the northern extent of white spruce appears to be limited by lack of warmth while the southern extent is controlled by lack of moisture (Brooks *et al.*, 1998).

A much greater area and the majority of biomass production of white spruce forest occurs in lower elevation stands (Ruess *et al.*, 1996) rather than marginal treeline stands. White spruce-dominated forest types make up about 12 million ha or 26% of the Alaska boreal forest, including 2.8 million ha or 51% of the commercially productive forest area (Labau and van Hees, 1990). Thus changes in the annual growth of low-elevation, productive white spruce forest types will be a major factor controlling variability of carbon dioxide uptake in the boreal forest of North America.

The negative relationship between the radial growth of productive white spruce on warm, low elevation sites in interior Alaska and summer temperatures is consistent throughout the entire 20^{th} century (Barber *et al.*, 2000). Given the limited annual precipitation in Interior Alaska, evapotranspiration as influenced primarily by high summer temperature becomes the limiting factor in annual radial growth. Both $\delta^{13}C$ discrimination and maximum latewood density properties of white spruce tree-rings, which represent a physiological signal of drought stress internal to the tree, are highly correlated with Fairbanks summer temperature during the 20^{th} century (Barber et al. 2000).

In this paper we examine ring-width, maximum late-wood density and carbon-13 discrimination of annual tree-rings of white spruce in interior Alaska to reconstruct and interpret 19th century climate. We use the relationship of combined $\delta^{13}C$ discrimination and maximum latewood density to mean summer temperature (May-August) to model summer temperature in the 19th century when climate records are not available. We also evaluate relative standardized radial growth of white spruce in the context of reconstructed climate.

Methods

All tree-rings in the sample used for this analysis are representative of white spruce in mature and old stands that are dominant on the contemporary landscape of interior Alaska, including trees across a broad range of diameters. We measured three properties of tree-rings, $^{13}C/^{12}C$ isotope ratios, maximum latewood density and ring-width. Trees at each site were crossdated using the software COEFECHA (Holmes, 2000) and visually by comparison of marker years to insure year-to-year correspondence in all measurements.

For development of ring-width chronologies, we used the 10 oldest stands (half of the sample) across interior Alaska from the calibration set used in Barber et al. (2000) (Figure 5.2). Each individual site was crossdated using COEFECHA and individual trees were corrected where possible if a problem was identified. Trees that didn't correlate with the master chronology ($r<0.4$) were excluded. Sample depth for individual trees during the period 1800 to 1996 varied between 43 and 220 (Figure 5.3). Ring-width was detrended using conservative negative exponential or straight-line curve fits (Fritts, 1976) and a chronology was created for each site using ARSTAN (Holmes, 2000). A master chronology for interior Alaska was then created using the 10 individual stand chronologies as the ARSTAN input. Sample depth of stands varied between 2 and 10 (Figure 5.3). Ring-width chronology values prior to 1816 had fewer than half of the stands or individual trees contributing to the chronology and reconstructions in those years should be treated with caution because of low sample depth (Figure 5.3). While sample depth also decreases markedly in the early 1990s, recorded data provide a check on the reconstruction through this period.

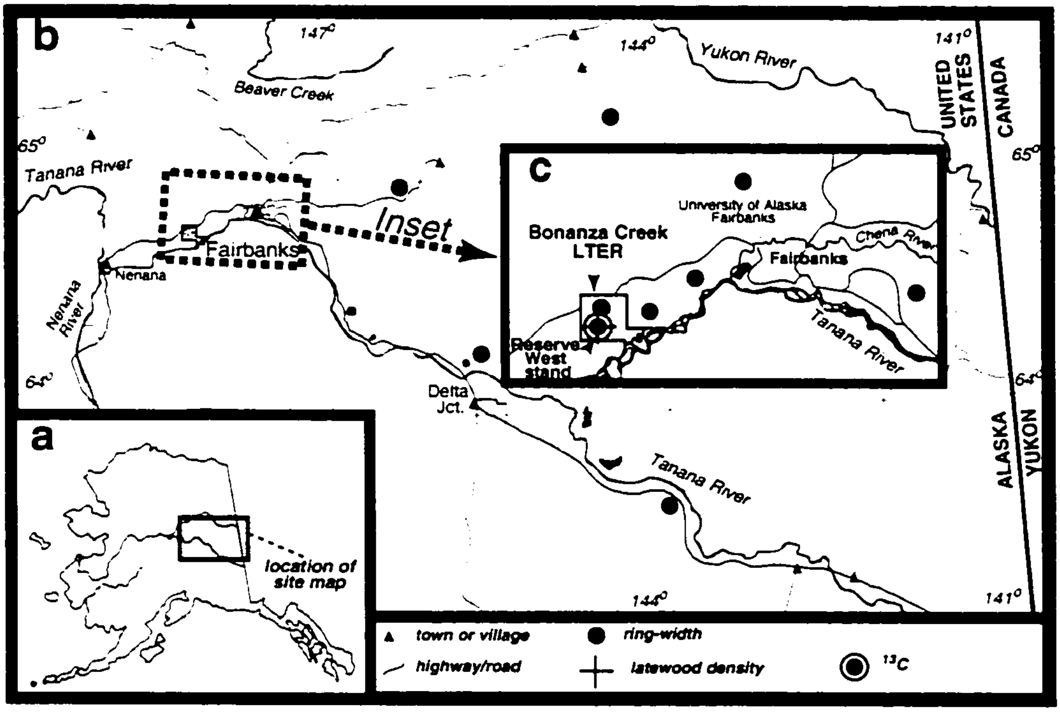

Figure 5.2. Location of tree-ring sampling sites in interior Alaska. Note the co-occurrence of ring-width, latewood density, and d13C sampling in the Reserve West stand at Bonanza Creek Long-Term Ecological Research (LTER) site.

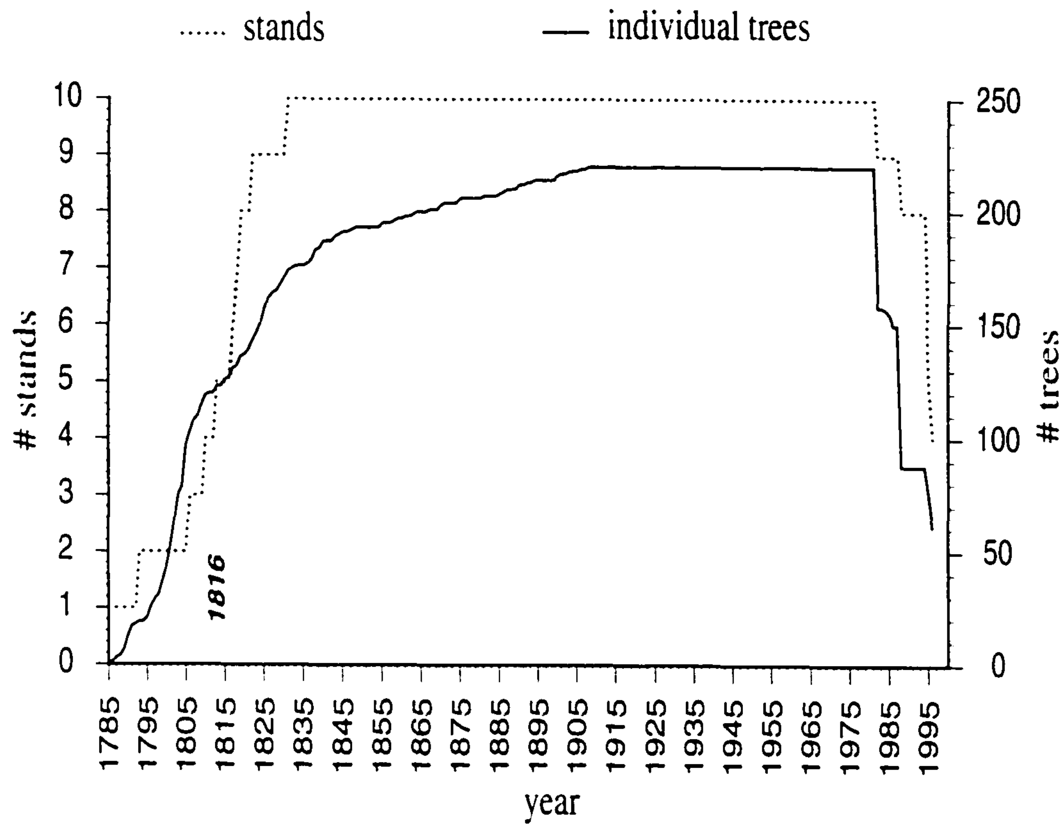

Figure 5.3. Sample depth for ring-width calculated as number of trees and number of stands. Note the fall-off in sample depth before 1816.

Stable carbon isotope ratios ($\delta^{13}C$) were measured only at the Reserve West stand. Growth patterns of dominant trees at Reserve West were highly correlated with the other 9 stands and the Fairbanks summer temperature record. Four orthogonal cores from each of 4 trees were pooled together give an isotope value representative of the site (Leavitt and Danzer, 1992). We obtained isotope measurements for the period 1800-1981 from 4 orthogonal wedges from each of 4 harvested stump disks of trees killed in a 1983 forest fire. For the period 1967-1996 isotope samples were obtained from 4 cores collected at each of 4 surviving trees near the fire perimeter. Four orthogonal samples of annual rings were excised, ground, and blended for each tree and then combined with the same year wood from the other trees. Hollocellulose was extracted from the annual wood (Leavitt and Danzer, 1992) and carbon isotope ratios were analyzed on a Europa 20/20 continuous flow mass spectrometer for the years 1800-1996. Values are reported relative to the PDB standard and the precision of the analyses was $\pm$ 0.1‰. The overlap period, 1967-1981, was used to determine site-specific isotopic offsets between the fire-killed and surviving trees. Isotope trends for the period of 1967-1981 follow similar curves from both sets of wood samples, but there is a slight offset (0.7‰) in the $\delta^{13}C$ of the fire-killed vs. the live trees, likely due to site-specific effects. A correction was made by adding this offset into the 1982-1996 data set to bring it into the same range as the longer data set. To correct for $\delta^{13}C$ changes in atmospheric CO_2 over the last 150 years due to fossil fuel and biomass combustion we used the Law Dome Antarctic ice core data (1.53‰ change over 150 years (Francey *et al.*, 1999)). Discrimination was calculated as the difference between the $\delta^{13}C$ of the atmosphere (ice core) minus the measure of $\delta^{13}C$ in the wood. Thus a continuous record of $\delta^{13}C$ discrimination was determined for 1800-1996. We confirmed the trends in $\delta^{13}C$ seen at Reserve West in limited samples covering 3 to 4 decades collected from 2 other sites located about 30 km from BNZ.

Maximum latewood density was measured by x-ray attenuation at Lamont-Doherty Earth Observatory. Tree cores were collected from 3 of the 10 sites (Dune Lake (64°25'N, 149°54'W), Jan Lake (63°34'N, 143°54'W) and the Reserve West stand at Bonanza Creek LTER (64°44'N, 148°18'W) for density analysis. A total of 14 cores from 12 trees were used from Reserve West, 19 cores from 9 trees for Jan Lake and 18 cores from 8 trees from Dune Lake. Density chronologies for individual trees from each site were run separately through ARSTAN and a master chronology was created for each site.

We determined relationships of monthly temperature and precipitation to standard chronologies of ring-width, $\delta^{13}C$ discrimination, and maximum density for growth year (Sept-Aug) and for year prior to growth (lag –1) using the principal components multiple regression software PCREG (Holmes, 2000).

Results

Standardized maximum latewood density chronologies of the 3 sampled sites created from the horizontal detrending were most highly correlated with average Fairbanks May-August temperatures (Table 5.1) in PCREG. Unfortunately, all but one of the Dune and Jan Lake tree cores were under 100 years old (Table 1), not long enough for the reconstruction of the 19th century, or lacked sufficient sample depth (Table 1).

Table 5.1. Statistics on the correlation of May-Aug temperature at Fairbanks (1906-1996) with standardized maximum latewood density at three individual sites.

Location	**r**	**P**	**N (1906-1996)**	**Total #Years**
Reserve West	0.707	.00001	13	215
Dune Lake	0.649	.00001	4-18	103
Jan Lake	0.547	.00001	5-17	190 (one core)

Table 5.2. Correlation coefficients between the three maximum latewood density sites.

Location	**Reserve West**	**Dune Lake**	**Jan Lake**
Reserve West	1.0	-	-
Dune Lake	0.567	1.0	-
Jan Lake	0.646	0.740	1.0

Since all three sites were highly correlated with each other (Table 5.2) and with the Fairbanks climate, only the Reserve West maximum latewood density chronology, which was over 200 years in length and of sufficient sample depth, was used for the reconstruction of Fairbanks summer temperatures.

All three tree-ring parameters, $\delta^{13}C$ discrimination, maximum latewood density and 10-stand ring-width, are highly correlated with mean summer temperature (May-

Aug) at Fairbanks for the contemporary growth year (Barber *et al.*, 2000). However, ring-width is the most autoregressive and is significantly correlated with summer temperature one and two years prior to the year in which the ring was formed (Barber *et al.*, 2000). Maximum latewood density and $\delta^{13}C$ discrimination are much less autocorrelated and are most highly correlated with summer temperature in the year of ring formation (Barber *et al.*, 2000).

The best reconstruction of summer (May through August) temperature produced by a single tree-ring parameter was produced by $\delta^{13}C$ discrimination, which explained 46.3% of the variance (Table 5.3). Both the calibration and verification periods produced a significant correlation (Table 5.3).

Table 5.3. Calibration-verification statistics for reconstruction of May-Aug temperature at Fairbanks based on ^{13}C discrimination.

Statistics	**1906-1996**	**1906-1950**	**1951-1996**
r	-0.680***	-0.576***	-0.648***
r2	0.463	0.332	0.420
Adj. r2	0.457	0.317	0.407
S		0.659***	0.546***
RE		0.569	0.573
CE		0.141	0.128
Sign test		29+ 17-	28+ 17-
Cross Product Means Test		0.443**	0.483**

r = multiple correlation coefficient for predictor 1 (^{13}Cdiscrim.)
r^2 = variance explained
adj. r^2 = r^2 adjusted for loss of degrees of freedom
S = Spearman rank correlation coefficient
RE = reduction of error statistic
CE = coefficient of efficiency
*Significant at the < 0.01 level
**Significant at the < 0.001 level

***Significant at the < 0.0001 level

Maximum latewood density also produced a summer temperature reconstruction with high, but slightly lower statistical significance, explaining 38.7% of the variance (Table 5.4). The reconstruction of summer temperature produced by ring-width alone was significant over the entire period of instrument record, explaining 29.6% of the overall variance, but it was not significant during the first half of the calibration period (1906-1950) (Table 5.5).

We chose $\delta^{13}C$ discrimination and maximum density as the independent (predictor) variables of summer temperature based on their higher predictive ability. These predictors were supplied to PCREG to reconstruct average May through August Fairbanks temperatures (Figure 5.4). In order to check temporal consistency, the period of recorded climate was divided in half, and each half was used separately for calibration and verification. The reconstruction based on combined $\delta^{13}C$ discrimination and maximum latewood density explained more of the overall variance (59.9%) and more of the variance in the calibration and verification periods than any of the individual ring parameters (Table 5.6). Adding ring-width to the reconstruction did not increase the explained variance (in fact it declined slightly), and therefore we used only $\delta^{13}C$ discrimination and maximum latewood density for the final reconstruction.

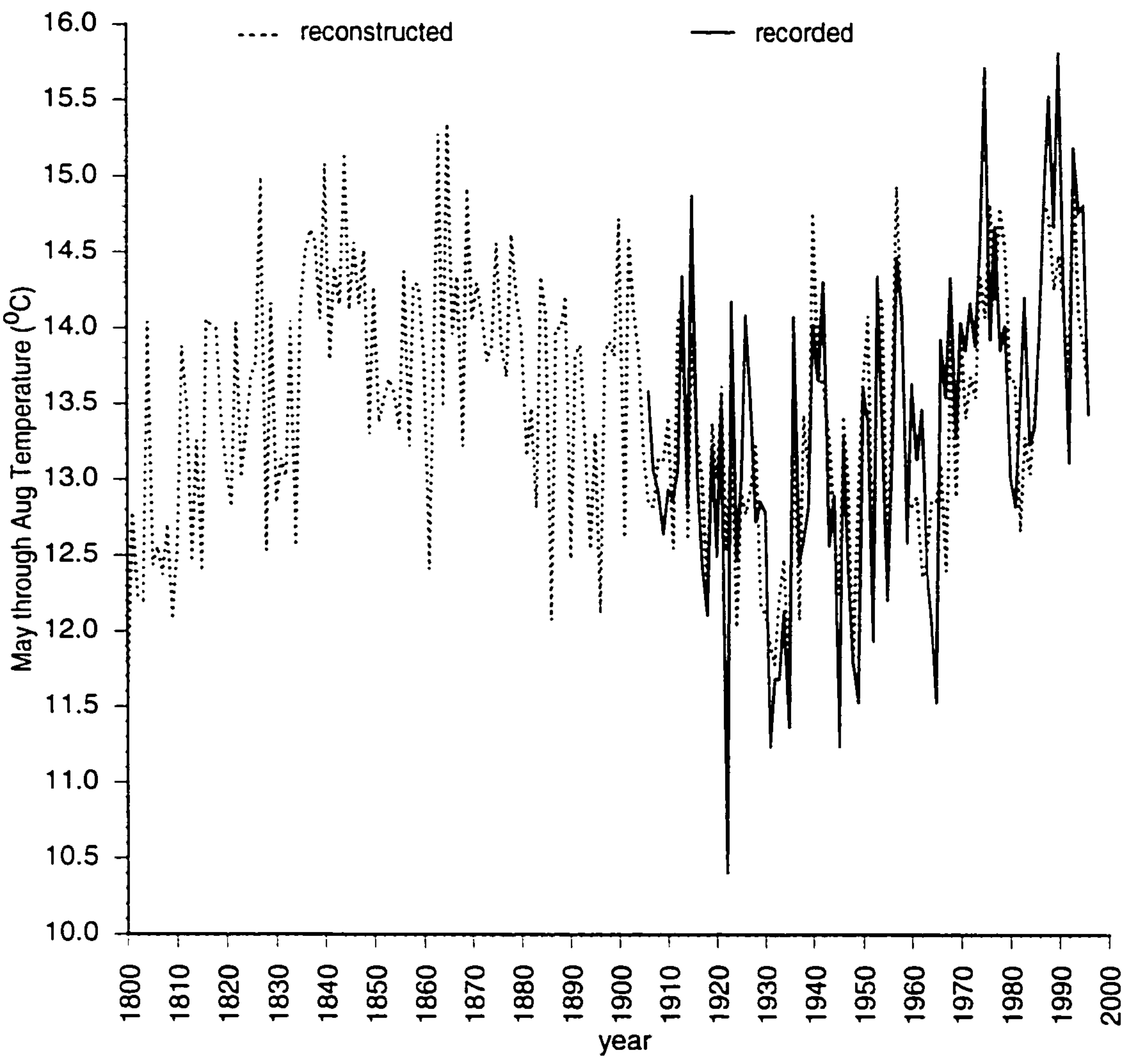

Figure 5.4. Reconstruction of May-August temperature for Fairbanks Alaska. Reconstruction is based on ^{13}C discrimination and maximum latewood density compared to recorded data.

Table 5.4. Calibration-verification statistics for reconstruction of May-Aug temperature at Fairbanks based on maximum latewood density.

Statistics	1906-1996	1906-1950	1951-1996
r	0.622***	0.540***	0.565***
r2	0.387	0.292	0.319
Adj. r2	0.380	0.275	0.304
S		0.547***	0.531**
RE		0.494	0.477
CE		-0.008	-0.082
Sign test		27+ 19-	29+ 15-
Cross Product Means Test		0.439***	0.208

Symbols as on Table 5.3.

For the earlier calibration period, the adjusted r^2 is 0.499. The RE (reduction of error statistic) is strongly positive (0.710) over the verification period (1951-1996), which shows that there is considerable skill in the verification estimates as compared to the calibration period mean. However, it may also partly reflect a difference between the means of the calibration and verification periods (D'Arrigo and Jacoby Jr., 1992). The coefficient of efficiency (CE) statistic differs from the RE, in that it compares the estimated data for the verification period to the mean of this period and this number is also strongly positive for both halves of the calibration/verification data. For additional verification, the Spearman rank correlation and product means test statistics were calculated and both were significant at the 0.0001 level. The cross-product means test measures the level of agreement between the actual and estimated values and takes into account the sign and magnitude of departures from the calibration average (Fritts, 1976).

Table 5.5. Calibration-verification statistics for reconstruction of May-Aug temperature at Fairbanks based on ring-width.

Statistics	1906-1996	1906-1950	1951-1996
r	-0.544***	-0.299*	-0.548***
r2	0.296	0.090	0.301
Adj. r2	0.288	0.068	0.285
S		0.495**	0.299*
RE		0.367	0.435
CE		-0.263	-0.155
Sign test		21+ 25-	28+ 17-
Product Means Test		0.355	0.171

Symbols as on Table 5.3.

When the calibration and verification periods are reversed, the adjusted r^2 is 0.511, the RE is 0.710, the CE is 0.408, the Spearman rank correlation is 0.727, and the cross product mean test is 0.499, all highly significant. These results suggest that the model used here (combined $\delta^{13}C$ discrimination and maximum latewood density) passes the critical tests for verification and is optimized by achieving the greatest predictive capability for the fewest independent variables.

5. 6. Calibration-verification statistics for reconstruction of May-Aug temperature at Fairbanks based on of maximum latewood density and ^{13}C discrimination.

Statistics	1906-1996	1906-1950	1951-1996
r_1	-0.680***	-0.576***	-0.648***
r_2	0.622***	0.540***	0.566***
r^2	0.599	0.511	0.522
Adj. r^2	0.595	0.499	0.511
S		0.719***	0.727*
RE		0.722	0.710
CE		0.445	0.408
Sign test		38+ 8-	33+ 12-
Cross Product Means Test		0.543***	0.499***

Symbols as on Table 5.3, with

r_1 = multiple correlation coefficient for predictor 1 (^{13}C discrimination)

r_2 = multiple correlation coefficient for predictor 2 (max density)

The temperature reconstruction for the 200-year time period shows a distinctive low resolution sinusoidal pattern with high-resolution decadal-scale periods overlain (Figure 5.4). These decadal-scale cycles shift rapidly over the course of a few years. The average reconstructed summer temperature for the entire period (1800-1996) is 13.49°C (Figure 5.4). The mean of both the 20th century (1906-1996) recorded temperature and the reconstructed temperature is 13.31°C.

Since the reconstruction showed distinctive decadal to multi-decadal periods, we divided it into regimes (Figure 5.5). In order to establish our divisions, we used a moving window analysis with squared Euclidean distance metrics (MW SED) (Johnson *et al.*, 1992; Turner *et al.*, 1991) on the May-August temperature reconstruction. We chose a 17-year period for the moving window because the data display a near 11-yr periodicity (Juday, 1984) and a period approximately half again as long (17 years) will optimize the

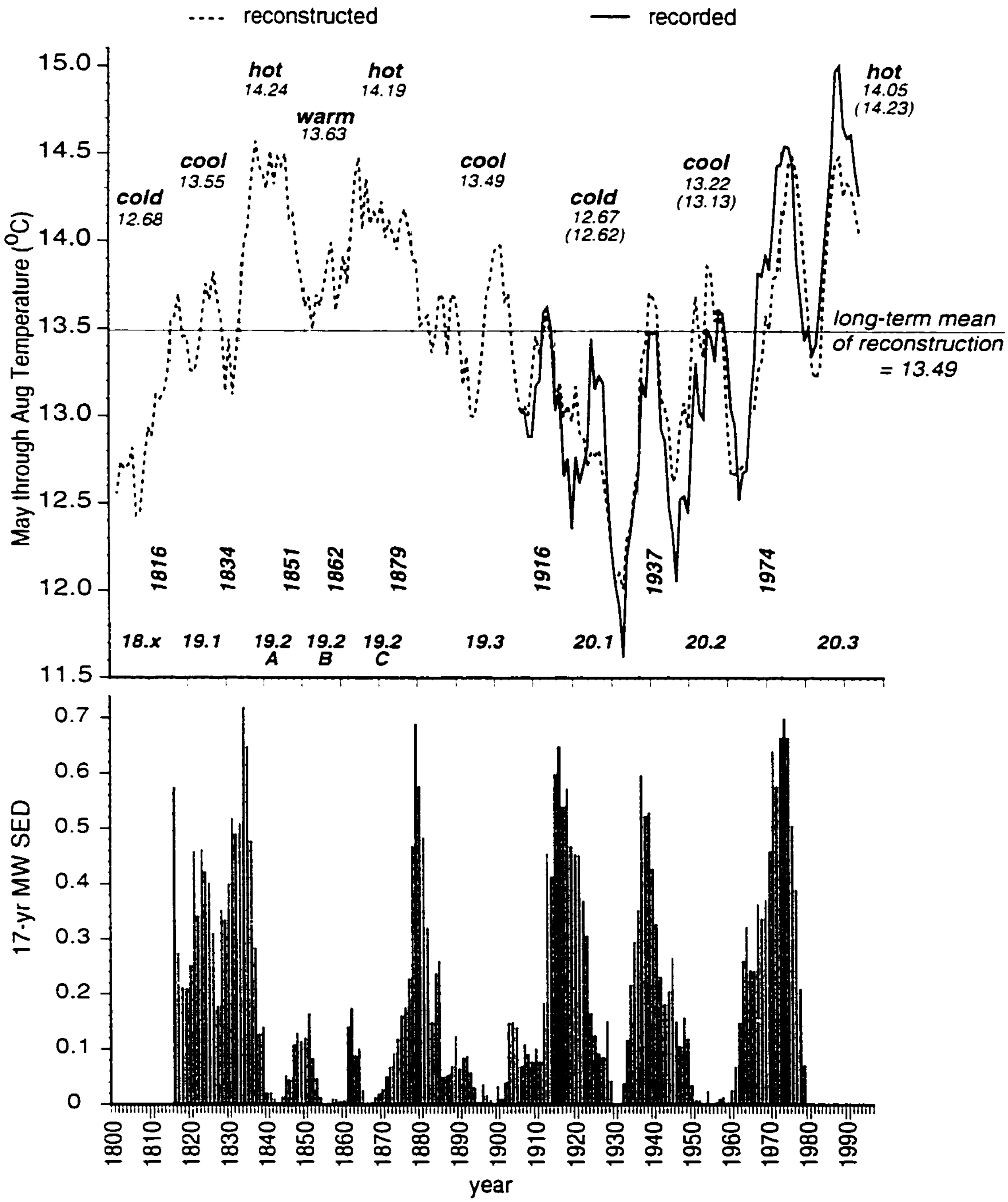

Figure 5.5. Reconstructed summer temperature divided into regimes. Graph includes dates of change (top), and 17-year moving-window squared Euclidean distance (MW SED) metric used to define nodes of change (bottom). Note identifying regime numbering system composed of century identifier (left of decimal) and sequential numeral within the century (right of decimal) along horizontal axis. Reconstructed mean temperature is displayed for each regime (recorded mean in parentheses).

definition of the decadal variability in the reconstruction. The results of the MW SED show peaks or spikes where the greatest change occurred, and we used these spikes to define the boundaries of climate regimes. Our proposed climate regimes are defined as multi-decadal periods sandwiched between periods of rapid climate changes of a few years in duration (Figure 5.5). For convenience, we number climate regimes for the century in which they began, and assign a decimal of 1 for the first regime initiated in the century and increasing decimal numbers for successive regimes of that century. Thus the first regime initiated in the 20^{th} century is labeled 20.1, etc. We have further divided one regime (19.2) into three sub-regimes (A, B, and C) of lesser magnitude change.

Based on these criteria, the 20^{th} century contains Regimes 20.1, 20.2, and 20.3, and the 19^{th} century is divided into 3 regimes. We tentatively identify a rapid climate change at about 1816 based on an apparent major change in variables compared to Regime 19.1, although we do not have MW SED values calculated prior to that year. These early years of the 19^{th} century appear to have been part of a regime that began in the 18^{Th} century and since we are unsure of the number of regimes contained in that century, we label that regime as 18.x.

We were unsuccessful in reconstructing precipitation. Although ring-width correlates well with an index of summer temperature and growth year precipitation (Barber *et al.*, 2000), attempts to isolate and reconstruct precipitation failed the test of significance in PCREG.

Discussion

The 200-year reconstruction based on the combination of tree-ring proxies $\delta^{13}C$ discrimination and maximum latewood density, shows excellent agreement with the recorded Fairbanks mean May through August temperatures (Figure 5.4). Recorded summer temperature displays greater amplitude than reconstructed temperature, which suggests that although $\delta^{13}C$ and density are superior to ring-width in reconstructing summer temperatures on an annual basis, they still retain a small amount of autoregression. Both the reconstructed and recorded summer temperatures of the latter part of the 20^{th} century, particularly from 1974 onward, are characterized by some of the warmest summers in the 200-year interval (Figure 5.5). The first half of the 20^{th} century is characterized by the coolest summers of the 200-year period of reconstruction.

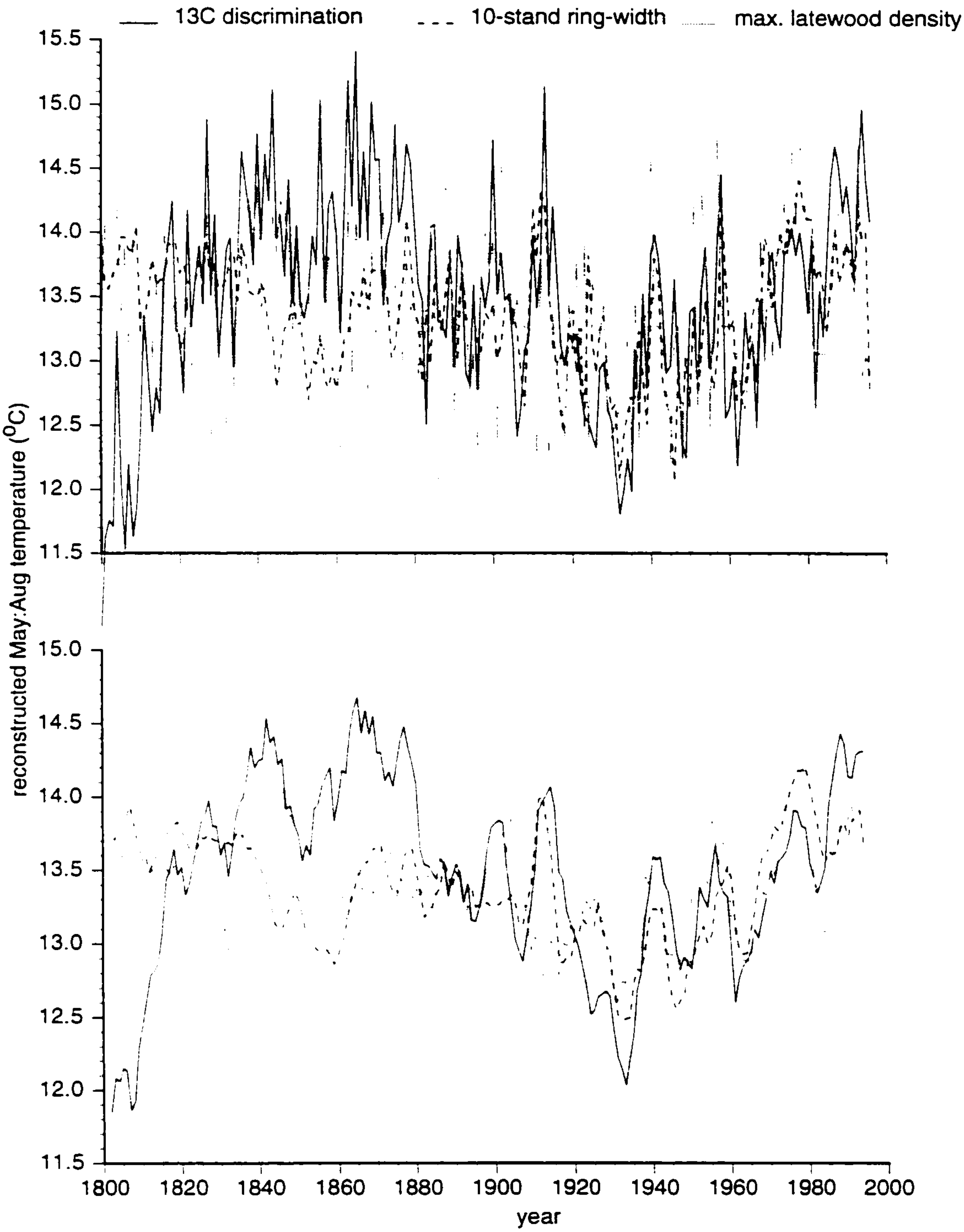

Figure 5.6. Three individual reconstructions of Fairbanks summer temperature. The reconstructions are based on the three individual proxies of ^{13}C discrimination, maximum latewood density, and ring-width index. Annual values (top), and 5-yr. running mean values (bottom).

Interestingly, mid-19th century summer temperatures reconstruct as some of the warmest over the 200-year period.

In an attempt to look more closely at our surprising reconstruction of summer warmth during the mid-19th century, we looked for discrepancies in the individual reconstructions of summer temperatures by the three proxies ($\delta^{13}C$ discrimination, maximum latewood density and ring-width index) (Figure 5.6). All three proxies by themselves reconstruct overall warmer summer temperatures in the 19th century than temperatures in the early to mid-20th century. There are some discrepancies between the three proxies, especially for the early part of the 19th century (1800-1815). The isotope reconstruction shows the greatest anomaly, reconstructing much warmer temperatures from the mid- 1850's to 1879. It also suggests cooler temperatures in the earliest decade of the 19th century than the other two proxies do. But all three reconstruct increasing temperatures around 1835, with either relatively warm or moderate temperatures through 1879 compared to regimes 20.1 and 20.2 (1916-1974). Thus, all three proxies agree on the interpretation of a fairly warm mid 19th century with cooling into the middle decades of the 20th century.

Most reconstructions of Northern Hemisphere temperature indicate cool annual temperatures in the earlier part of the 19th century, which is generally attributed to the latter part of the Little Ice Age (Bradley and Jones, 1993). Published reconstructions for the Northern Hemisphere indicate a warming in the mid-19th century that continued into the early 20th century (Jacoby and D'Arrigo, 1989; Overpeck *et al.*, 1997). By contrast our data show that the coolest part of the 200-year reconstructed summer temperature for interior Alaska was the early part of the 20th century (1916-1937), a cool interval not seen in most Northern Hemisphere temperature reconstructions or records (Jacoby and D'Arrigo, 1989; Mann *et al.*, 1998; Overpeck *et al.*, 1997). Our results suggest that in fact, interior Alaska may have experienced temperature trends different from the overall trend in northern North America during the period of analysis. For example, Northern Hemisphere mean temperatures (Mann *et al.*, 1998) (Figure 5.7) show a fairly steady increase in warm season temperatures anomalies from the earliest 20th century until the mid-1950's, while Fairbanks recorded mean summer temperature shows a cooling from around 1916 through 1937 (regime 20.1). The Fairbanks warm season record displays a brief warming in the early 1940's and at 1957-58. It then remains generally cool until

1974, after which temperatures increased up to the present. The Northern Hemisphere warm season temperature anomaly shows little trend from the mid-1950's to about the mid-1980's at which point trends in the two records come into agreement and show warming to the highest levels of the century (Figure 5.7). The discrepancies between the two temperature records may be indicative of differences in atmospheric circulation patterns in the earlier part of the 20^{th} century, and/or common forcing from greenhouse gases in the latter part of the 20^{th} century.

Additional data indicating unusual summer warmth in central Alaska in the mid-19^{th} century adds support to our reconstruction. Gridded temperature patterns in the Northern Hemisphere (Mann *et al.*, 1998) based on multiproxy EOF reconstructions reveal anomalies in Alaska temperatures as compared with the rest of northern North America. One interesting result from that study shows that around 1834 (at which time our reconstruction shows warming in interior Alaska summer temperatures), most of northern North America was very cold, but Alaska was warmer than average. The following historical account from 1822 given by Nordkvist from the Bering Sea region hints at relatively warm conditions in Alaska in 1822: "Not only in the summer, but in the winter the ocean was free of ice sometimes with a wide strip of water up to at least 200 miles away from the shore" (Koskey and Yamin, 2001). Thus our reconstruction may be consistent with an atmospheric circulation pattern that resulted in a very localized regionally warmer climate for Alaska, but a cooler climate elsewhere over much of northern North America.

Two alternative hypotheses may account for our reconstruction of warm temperatures in the first half of the 19th century:

1) The proxy signal ($\delta^{13}C$ discrimination and maximum latewood density) upon which the reconstruction is based may have been influenced by properties of the site or trees that were unique to the early life of the stand (juvenile effect), a time that happens to correspond to the early to mid-19th century. Once the trees or stand matured, these effects disappeared and the period of reconstruction overlap with recorded climate data essentially represents a different calibration interval.

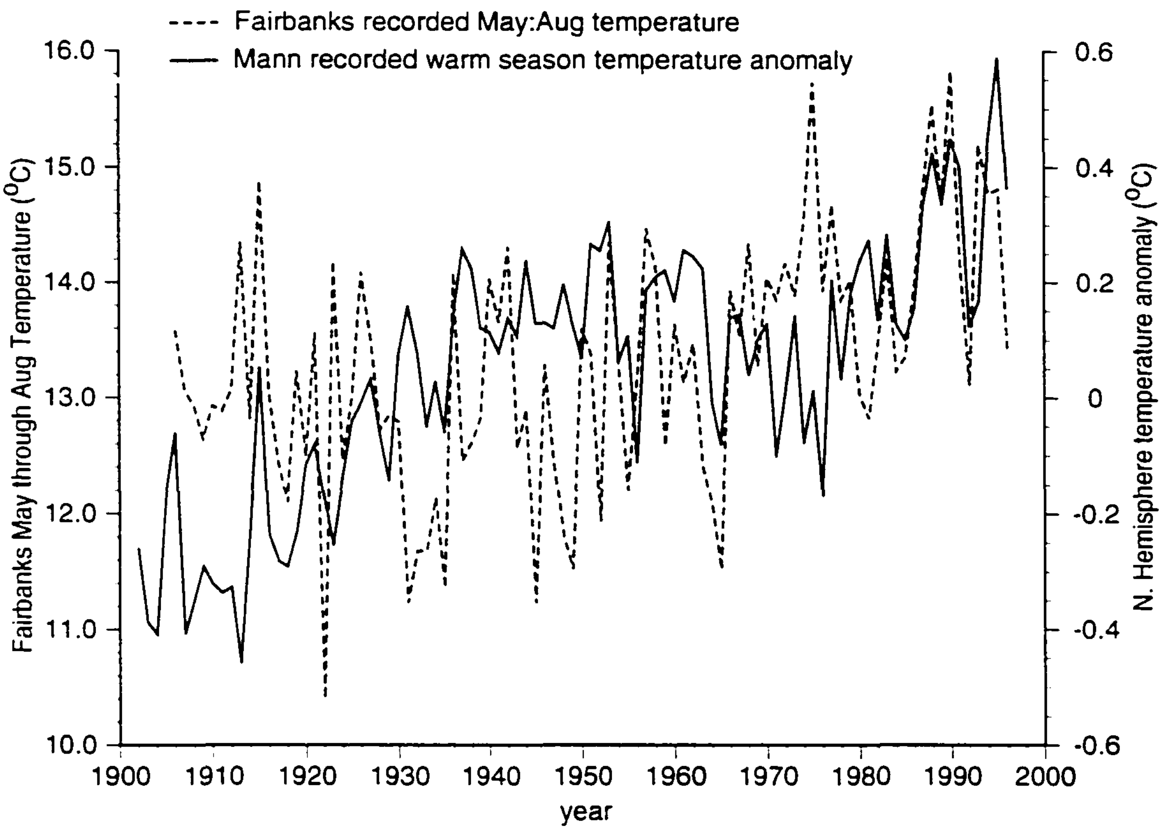

Figure 5.7. Fairbanks mean summer temperature vs. Mann et al. (1998) Northern Hemisphere warm season temperature anomaly. Scales have been adjusted to produce ranges that provide maximum overlap of lines.

2) Conditions with no modern analog may have existed in the 19th century. There are 2 parts to this hypothesis:

a) The early 19th century may have been colder than the range of temperature that occurred in the 20th century and consequently the white spruce trees on the sites we sampled may have been limited by the direct effect of very low temperatures. This explanation would require that an ecological threshold be crossed so that the trees changed their growth response to temperature from positive (19th century) to negative (20th century).

b) The 19th century may have been moderately cool to cold, but very much drier than the 20th century. There are no sustained periods of cold and dry summer conditions in the 20th century record as seen earlier in this paper. We demonstrated that drought stress induced by high temperatures was the limiting factor behind reduced radial growth in white spruce on productive sites in central Alaska in the 20th century (Barber *et al.*, 2000). Extremely arid conditions in the 19th century could have produced physiological drought in the trees directly (without the influence of high temperatures) and limited their growth, even during a period of cooler summer temperatures than during the 20th century.

In support of hypothesis 1, there may be a juvenile effect which shows up in the isotope data (Bert *et al.*, 1997) which would have driven discrimination down as the trees grew and the canopy expanded and then closed. Autogenic effects on a site are prominent early in the life of a forest and are known to influence moisture stress (Bert *et al.*, 1997). The trees used for $\delta^{13}C$ analysis came from a stand that regenerated around 1785. As the trees grew in the first several decades of the life of the stand, there may have been an increase in moisture stress due primarily to the development of a full canopy. In particular, the higher levels of $\delta^{13}C$ discrimination in the first decades of the 19th century might be influenced by the lower levels of evapotranspiration in a young stand of trees compared to the same stand at canopy closure. However, autogenic effects on the site would, if anything, be greater in the 20th century when evapotranspiring leaf area reached even greater levels than in the first few decades in the life of the stand. So while decreasing discrimination suggests increasing moisture stress which may have been partly a result of less soil moisture available to the trees as a result of their own growth, there is no reason to believe that the effect was uniquely strong in the 19th century. In addition, the amplitude of change in $\delta^{13}C$ discrimination appears to be greater than could be produced by the unique growth characteristics of a particular age class of trees.

In support of hypothesis 2a, it is possible that during the first half of the 19th century (Regime 19.2), summers were so cold that a different growth response occurred in white spruce in interior Alaska. Other studies show either reduced sensitivity (Briffa *et al.*, 1998) or altered sensitivity (Jacoby and D'Arrigo, 1995) of boreal trees to temperature in recent decades. Whereas other studies (at treeline) show a positive response of radial growth to temperature, our trees show a negative response. It may be that low elevation trees with a negative response to temperature at the warmer end of the temperature continuum would have a positive response to temperature at the colder end, despite the fact that the negative growth response to summer temperature is linear within the range of modern temperature and precipitation. But given the consistent response of the three white spruce tree-ring properties across a wide range of climate conditions in this study, it seems unlikely that the fundamental relationship of these properties to temperature changed during the 19th century, although we don't know where a threshold, if any, lies. A possible method to resolve this issue would be to assemble proxy data of greater time depth from other living organisms with different ecological thresholds or in adjacent regions that do not experience as severe moisture limitations.

In support of hypothesis 2b, it may have been cool to cold during the early to mid-19th century in interior Alaska, but so dry that effective moisture (precipitation minus evaporation) was extremely low and therefore growth was limited. There is no modern analog in the recorded data for this type of climate (cool/dry). The two predominant climate modes in the 20th century recorded Fairbanks data are consistent with changes in overall circulation patterns that produce two prevalent anomalous summer climate patterns in interior Alaska: strong maritime (cool and wet) versus continental (hot and dry) conditions (Mock *et al.*, 1998). There is no period in the nearly 100 year Fairbanks record of sustained cool and dry conditions. It is also extremely difficult to find proxy data in Alaska to resolve fine scale temporal and spatial precipitation patterns. Achieving a pure precipitation signal from any of the few tree species in Alaska remains elusive.

While these alternative hypotheses represent possible explanations for our results, the weight of the evidence suggests that it is reasonable to conclude that the data represent some real differences between interior Alaska and the rest of northern North America.

A further comparison of other northwestern North America temperature reconstructions could shed additional light on peculiar features of the interior Alaska reconstruction we present here. Most temperature reconstructions are based on tree-ring proxies from treeline trees, where the environment is substantially different than the environment of the trees used in this study. A typical northwestern North America temperature reconstruction of mean annual temperature from treeline trees (Jacoby and D'Arrigo, 1989), compared with our reconstruction of summer mean temperature (Figure 5.8), shows that the two curves display a consistent but inverse relationship. However, the northwestern North America reconstruction generally shows the same nodes of change as defined in our system of regimes and displays the same prominent decadal-scale changes. This inverse relationship between the two records is puzzling although the consistency of the relationship between the two curves suggests that there were some real and synchronous changes in the environment over this time period.

The climate during regimes 20.1 and 20.2 was so favorable that these years represent the period of greatest relative radial growth of interior Alaska white spruce over the 200-year record by a substantial margin. A similarly large relative radial growth anomaly occurred at the same time at site 412[1] (67° 56'N, 162° 18'W) located in northwestern Alaska (Figure 5.9). Because summer temperature is an excellent predictor of radial growth of white spruce on productive sites across a broad area of central Alaska (Barber *et al.*, 2000), carbon uptake by this system almost certainly varies in accordance with our defined regimes and has diminished considerably in the current regime.

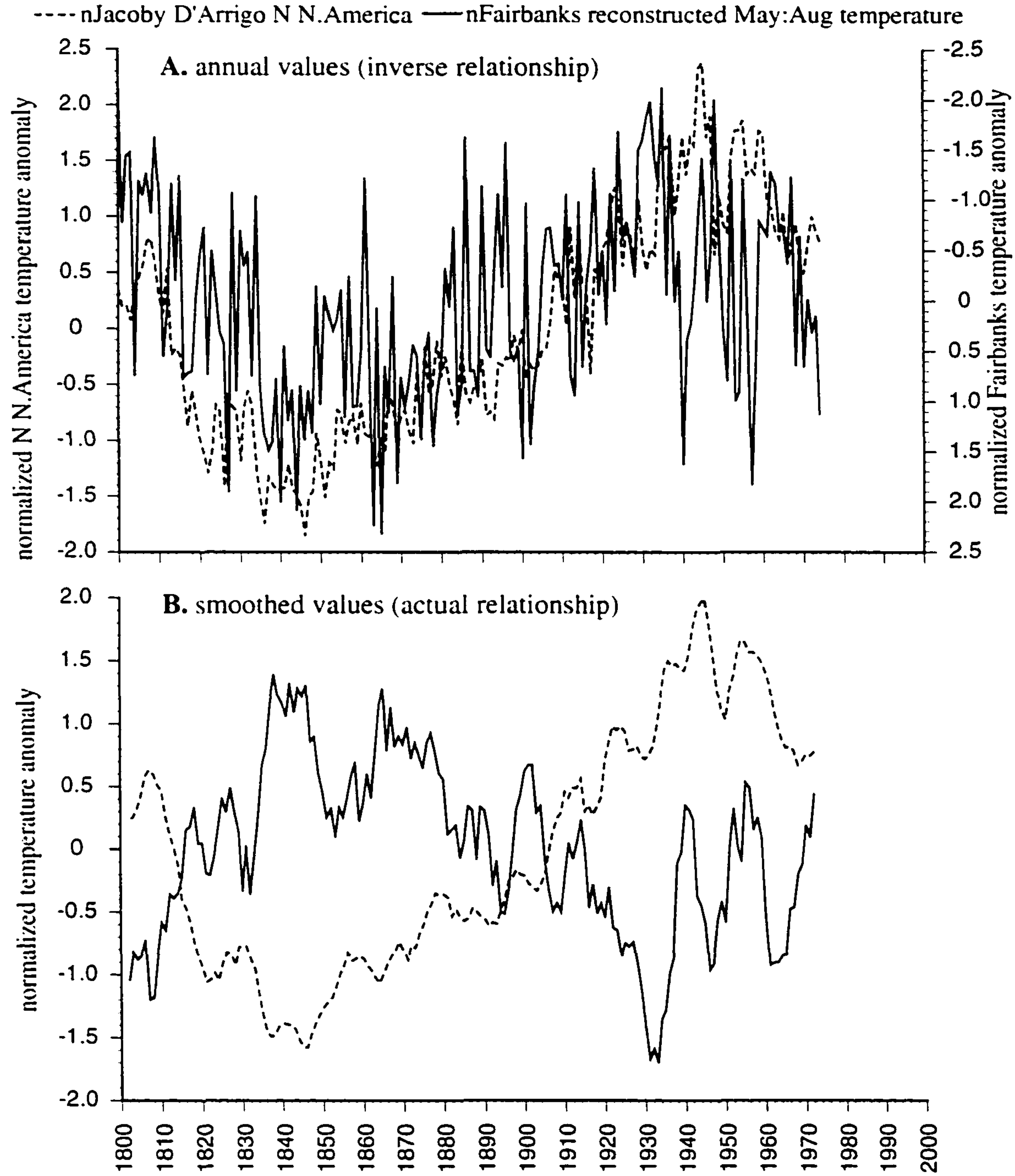

Figure 5.8. Northern North America mean annual temperature anomaly reconstruction versus Fairbanks reconstructed summer temperature anomaly. Northern North Americareconstruction is from trees at treeline (Jacoby and D'Arrigo, 1989) (A) Annual values with inversion of Fairbanks anomaly (B) Smoothed (5-yr. running mean) values (no inversion).

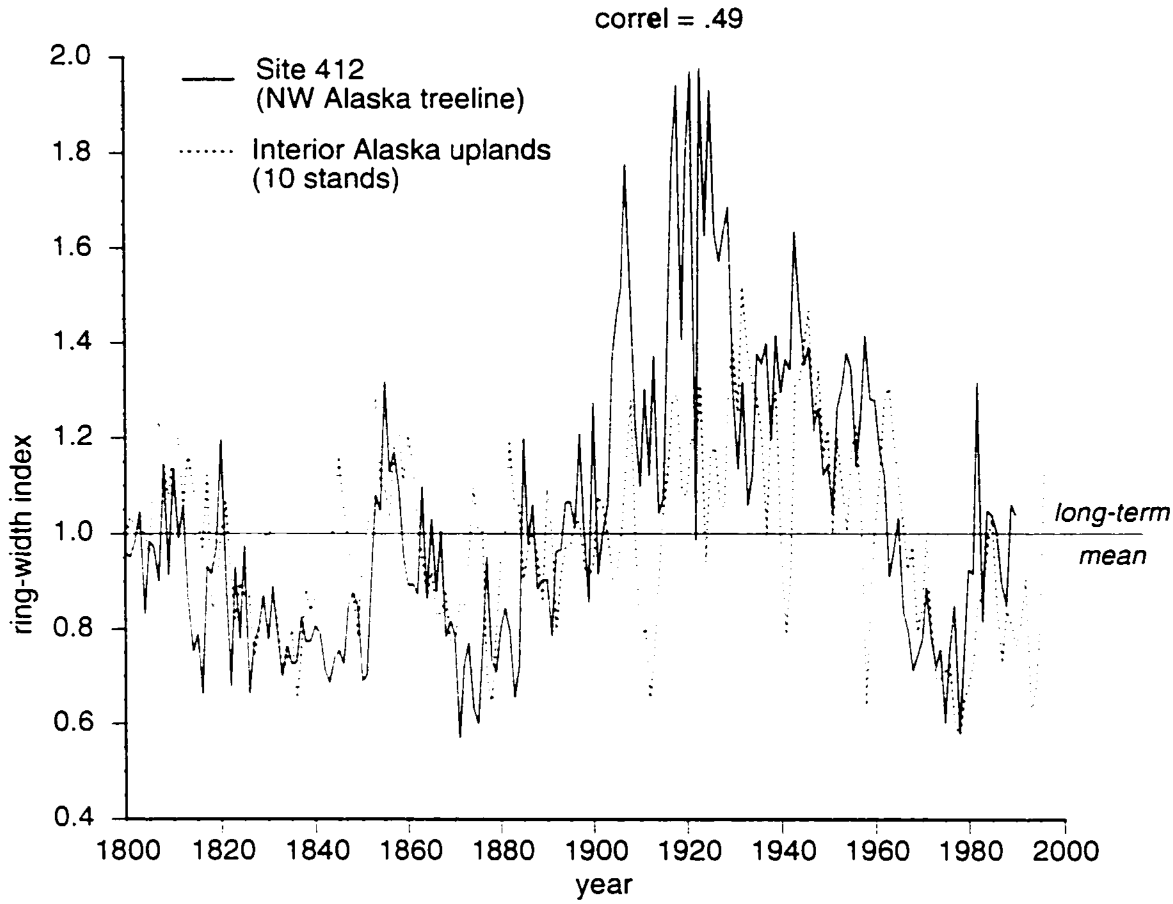

Figure 5.9. Comparison of relative radial growth of white spruce since 1800 from interior Alaska (productive forest) versus Northwestern Alaska (treeline).Values represent detrended, normalized transformation of ring-width, with mean set to 1.0. Pearson correlation = 0.49.

Conclusions

Our 200-year summer temperature reconstruction is one of the first regional scale high-resolution reconstructions for interior Alaska and shows excellent agreement with the recorded data from the 20th century. The reconstruction, as well as the recorded data, displays decadal-scale regimes, which appear to be the result of pervasive synoptic scale climate patterns. The early 20th century cool/wet period was conducive to the highest radial growth and carbon uptake by productive white spruce stands in interior Alaska over a 200-year period.

Our 200-yr reconstruction of summer temperatures in interior Alaska is anomalous from other Northern Hemisphere reconstructions for this time period. We cannot completely rule out autogenic factors playing some role in our reconstruction or that the sensitivity of tree-ring properties may have been different in the 19th century, although this seems unlikely. It could be that summers were cool to cold during much of then 19th century, while simultaneously so dry that moisture was limiting. But the evidence suggests that the anomalies we have identified may be real and reflect very different regional synoptic conditions between interior Alaska and the rest of northern North America. Alaska, as a peninsular extension of North America to the west, is peculiarly susceptible to the influx of marine air moving north off the Bering Sea and North Pacific. When this impulse is strong during the warm season, summer in central Alaska is anomalously cool and moist. When atmospheric circulation is dominated by blocking high pressure centered on central Alaska, the long days near the summer solstice cause hot and dry conditions. Alternation between these two modes of atmospheric circulation dominated the 20th century. Cool/wet conditions dominated from the early century until a shift to warm/dry conditions in the 1970's. This change could be interpreted either as a shift in atmospheric circulation to more frequent warm/dry types, or as decadal-scale cycles of the two predominant circulation types superimposed on a global-warming signal.

The summer warming evident since the regime shift around 1974 is unprecedented over the nearly100-year period of recorded data. Although our results suggest that synoptic conditions similar to the late 20th century warm period may have existed in the early to mid-19th century, additional evidence in the form of other proxies and examination of temperatures in different high latitude regions at synchronous periods may be able to further clarify the occurrence of past warm season temperature anomalies that we have identified in central Alaska.

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